Reconstructing the glacial history of Upgang, North Yorkshire: A North Sea Ice Lobe readvance during the termination of the Last Glacial Maximum?

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Abstract

Around Great Britain at the end of the Last Glacial Maximum (LGM) there is evidence for a number of glacial readvances, specifically in the Irish Sea Basin and Eastern Scotland. These readvances have been correlated with Heinrich Event 1 (H1) (16.1 – 17.6 cal. years BP; Bond *et al.* 1992). However, the presence of a synchronous H1 readvance on the Yorkshire coast has not been investigated, although it is known from previous work that a number of readvances occurred during the end of the LGM.

Therefore, this study aims to establish, firstly, if a readvance signal exists in Yorkshire and, secondly, if this signal can be correlated with H1. In order to achieve this, the coastal section of Upgang in North Yorkshire is examined using a multi-proxy approach. This involves the study of the sediments, clast form, clast fabric and clast lithological analysis along with geochemistry and geomorphological mapping. The sequence at Upgang shows a distinct advance – retreat – readvance signal. Two lower subglacial tills, D1 and D2 represent an initial ice advance and are a glaciotectonite or overrun proglacial thrust feature and a subglacial traction till respectively. Deposited above these is an extensive sand and gravels lithofacies association, indicative of an infilled lake after the ice had retreated. This lithofacies also represents the increasingly ice proximal nature of the site with the upper facies representing proglacial subaerial sandur sedimentation. The section is capped by the readvance till, D3, another subglacial traction till.

Using local vegetation and varve records from Kildale and the Tees Estuary a proxy climate record is produced. This along with the correlation of the Upgang tills with those at Dimlington allows the production of a chronostratigraphic framework. This shows that the initial ice advance occurred shortly after 21 000 cal. years BP. The readvance then occurred around 16 000 cal. years BP in conjunction with a deterioration in climate. Therefore, this readvance is tentatively correlated with Heinrich Event 1, which is also seen as the mechanism for other Last Glacial Maximum readvances around Great Britain, although without absolute dating control no great certainty can be placed upon this.

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1. Introduction

1.1 Rationale

The Last Glacial Maximum (LGM) occurred in Britain between 19 000 and 23 000 cal. years BP (Bowen *et al.* 2002). During this glacial episode the British-Irish Ice Sheet (BIIS) extended as far south as the Isles of Scilly in the Irish Sea Basin (ISB) (Scourse and Furze 2001, O'Cofaigh and Evans 2007) and North Norfolk on the east coast (Balson and Jeffrey 1991). In Britain the LGM is also referred to as the Dimlington Stadial (Lowe and Walker 1997). Throughout this thesis the term LGM will refer to the British Last Glacial Maximum and not the global LGM. Deglaciation of the BIS from the LGM occurred between 20 000 cal. years BP (Penny *et al.* 1969) and 13 000 cal. years BP (Beckett *et al.* 1981). These dates represent the termination of the LGM taken from the oldest date indicating glaciation in Yorkshire from Dimlington and the youngest date indicating an ice free area from Roos Bog.

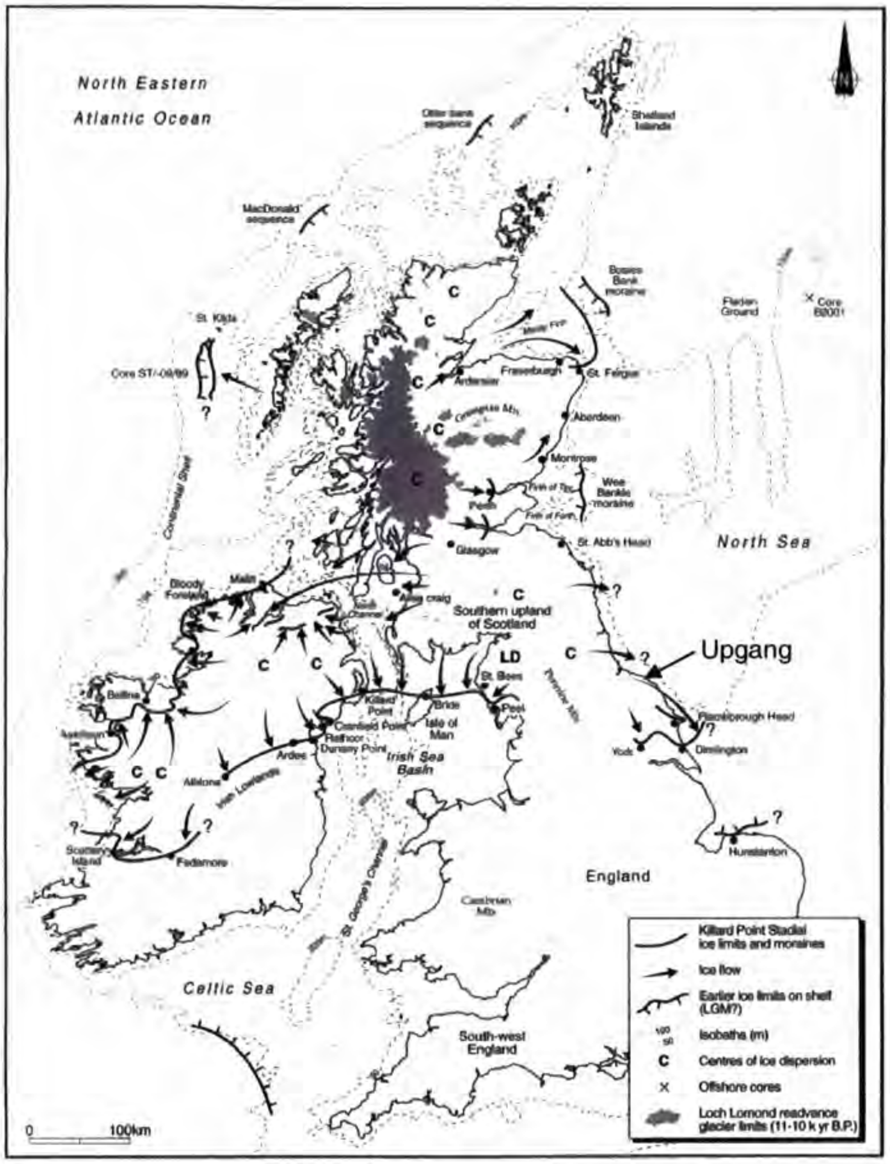


Figure 1: Map of Heinrich 1 readvance limits in Great Britain and Ireland. (McCabe *et al.* 1998).

During deglaciation from this episode large parts of Britain appear to show evidence for glacial readvances (E.g. Trotter 1929, Huddart 1971, Hall and Jarvis 1989, McCabe 1996). In 1998 McCabe *et al.* correlated one of these readvance in the ISB to Heinrich Event 1 (H1), 16.1 – 17.6 cal. years BP (Bond *et al.* 1992), sparking renewed interest in these LGM readvance signals. H1 represents a sub-millennial cooling phase caused by a high discharge of Laurentide Ice Sheet icebergs into the North Atlantic resulting in the shut down or weakening of the Thermohaline Circulation System (Lowe and Walker, 1997). McCabe *et al.* (1998) published sedimentological evidence from Killard Point that showed the existence of an ice sheet advance radiocarbon dated to c. 16 500 cal. years BP.

Following this further sedimentological evidence emerged suggesting a synchronous readvance of ice across the Irish Sea Basin correlating with H1 (Merritt and Auton 2000, Thomas *et al.* 2004, McCabe *et al.* 2004, Roberts *et al.* 2006, McCabe *et al.* 2007b). Evidence of an LGM readvance prior to the Loch Lomond Stadial (11 000 -10 000 cal. years BP (Lowe and Walker 1997)) has also been suggested on the east coast of Scotland (Merritt *et al.* 1995, Peacock 1997, McCabe *et al.* 2007a). However, many readvance signals across the country appear to be time transgressive due to the variance found in various dates, and this is a fundamental flaw in correlating these advances around Great Britain and Ireland and also with H1.

On the east coast of Britain any correlation between deglacial readvances and H1 has yet to be investigated (Teasdale and Hughes 1999). The timing of the glaciation of Yorkshire remains poorly researched. Radiocarbon dates taken from Dimlington of 20 931 – 22 281 cal. years BP and 20 559 – 19 258 cal. years BP (Penny *et al.* 1969) are of great importance. Taken from the Dimlington silts underlying the two tills of the Yorkshire coast they remain two of the few dates from the east coast constraining the timing of glaciation in the area. In addition, these dates were used by Rose (1985) to establish the climatostratigraphic horizon named the Dimlington Stadial which was seen to represent the main glacial episode of the Late Devensian in Britain. Deglacial dates for Yorkshire are taken from Kildale (19 144 - 20 537 cal. years BP (Jones 1977)) and Roos Bog (14 360 - 12 628 cal. years BP (Beckett *et al.* 1981)).

The east coast of Britain, including Yorkshire, is believed to have been glaciated by a dynamic ice lobe (Boulton and Hagdorn 2006) extending into the North Sea and along the coast. This lobe is thought to have been sourced from the major ice dispersal zones of Southern Scotland and the Lake District. Due to the dynamic and complex nature of the North Sea Ice Lobe a considerable amount of research has focused on the exposed glacial sediments of the Yorkshire coast, leading to numerous debates concerning the glacial history of the area. These have centred on the multiple till sequences relating to the LGM and pre LGM glaciations, along with the nature of the depositional processes are associated with these tills.

Since Madgett and Catt (1978), the glacial sequence of Basement-Skipsea-Withernsea till has been accepted as the established glacial stratigraphy for the Yorkshire coast. The Basement till was thought to have been deposited by a pre LGM glaciation while the deposition of the Skipsea and Withernsea tills have been dated to post 20 931 cal. years BP (Penny *et al.* 1969) as they overlie the dated Dimlington silts.

While the presence of three tills along the Holderness coast has largely been accepted, the means of deposition and ages of these sediments has been contested. Madgett and Catt (1978) suggested that the Skipsea and Withernsea tills had been deposited by a multilayered glacier. A multilayered glacier was classified by Catt (1991b p188) as 'a composite ice sheet comprising superimposed tributary glaciers'. However, Foster (1987) proposed that the Skipsea till had been deposited by ice from the Pennines and Lake District, before it was replaced by ice from Scotland which deposited the Withernsea till. Eyles *et al.* (1994) alternatively proposed that all the tills found along the Yorkshire coast were deposited by a surging glacier from the North Sea. This chronology was determined by the discovery of shell fragments with MIS 2 amino acid dates in the upper part of the Basement Till. Following on from this proposal, Evans *et al.* (1995) suggested that the tills had been deposited by vertical accretion with associated channel and cavity fills beneath an active ice sheet.

These debates concerning the glacial and process history of Yorkshire along with McCabe *et al.*(1998) questioning the validity of the Dimlington date, provide a framework with which to test the LGM and deglacial history of the North Yorkshire coast. By focussing on the relatively unstudied field site of Upgang, knowledge of the last glaciation of Yorkshire will be expanded upon. The establishment of any readvance signal along this coast, and any correlation to H1, will add to the growing number of sites within Great Britain where LGM readvances are found. This information can be used to improve glaciological models focusing on the last deglaciation (E.g. Lambeck 1995, 1996, Peltier *et al.* 2002, Boulton and Hagdorn 2006) which will in turn benefit the modelling of sea level and future climate change (Shennan and Horton 2002, Milne *et al.* 2006, IPCC 2001).

Therefore, this research seeks to establish the glacial history and mode of deposition of the sediments at Upgang, while attempting to establish if multiple ice flow phases and a potential LGM readvance signal is present. In addition, it aims to establish if these sediments can be linked to climatic forcing and a regional response by the BIIS to H1.

1.2 Aim

To establish whether there is a Last Glacial Maximum advance and readvance signal along the north east coast of England by reconstructing the glacial history of Upgang, North Yorkshire, and to test if this can be correlated to Heinrich Event 1.

1.3 Objectives

- To establish the sedimentological and structural properties of glaciogenic deposits at Upgang in order to reconstruct the depositional history of the site and local ice flow directions.
- To establish the provenance of the glaciogenic sediment in order to elaborate upon glacial transport pathways in the LGM BIIS.
- To establish a chronostratigraphical framework using regional vegetation records and other glacial stratigraphies (e.g. Dimlington and Kildale).
- To establish the glacial history of Upgang within the context of the LGM in Britain.

1.4 Hypotheses

1. The lower till is LGM in age and relates to an ice sheet advance.
2. The middle sands and gravels are associated with proglacial, subaerial sandur sedimentation processes.
3. The upper till is LGM in age relating to a readvance during deglaciation.
4. Both glacial units at Upgang correspond to post 20 000 cal. years BP advances.
5. The upper glacial advance correlates with the climatic cooling associated with H1

2. Background and Literature Review

2.1 Conceptual and theoretical frameworks.

Ice-ocean-climate science provides the framework for the study of ice sheet behaviour in relation to changing climate. Recent work looking at ice sheet readvance events around Great Britain and Ireland has focussed on the correlation between ice sheet behaviour and Heinrich Events (Heinrich, 1988), with Heinrich Event 1 (H1) in particular, having been linked to BIIS re-growth during deglaciation (McCabe *et al.* 1998, 2005, 2007; Merritt and Auton, 2000; Thomas *et al.* 2004)

Such studies have applied both a glacial landsystems and an event stratigraphy approach in an attempt to understand ice sheet – ocean – climate coupling. McCabe *et al.* (1996, 1998; 2005) were the first to combine glacial sedimentological analysis with the AMS radiocarbon dating of *in situ* forams to provide an event stratigraphy for ice sheet re-advance in the ISB during H1, and further inferred that a distinct glacial landsystem (drumlin and end moraine formation) was a product of this regional event. Knight (2003) and Clark *et al.* (2004) developed this approach further and matched evidence from Ireland with the climate signal from the North Atlantic Ocean and Greenland ice cores, and suggested global forcing mechanisms were behind BIIS fluctuations in the ISB. Other work has highlighted the complexity of the glacial landsystem response to BIIS re-advance during deglaciation, with complex ice marginal landsystems developing along the Cumbrian and Manx coastlines (Merritt and Auton, 2000, Thomas *et al.* 2004; Roberts *et al.* 2007).

2.2 Heinrich 1 literature

Bowen *et al.* (2002) suggested that the major advances of the BIIS were synchronous with the ocean-climate Heinrich events. Within this paper the LGM was dated to correspond with H2, with margin readvance signals correlating with H1.

H1, the most recent H event, was dated to approximately 14 000 ^{14}C years BP (16 400 cal. years BP) (McCabe and Clark 1998). This event forced the area of North Atlantic Deep Water (NADW) formation southwards, as far south as the Iberian Peninsula, decreasing sea surface temperatures (Prange *et al.* 2004). The position of the polar front was also altered, resulting in a drop in temperatures around the ocean margins (Clark *et al.* 2004b). Cruxifix and Berger (2002) modelled these changes associated with H1. The model indicated a drastic reduction in the Thermohaline conveyor at the start of H1 resulting in a cooling of the Northern Hemisphere. They showed that H1 caused the Thermohaline Conveyor to slow to the point of collapse between 18 000 and 14 000 years BP. This was linked with a limited cooling in Greenland and a more significant reduction in temperatures in the North Atlantic. Circulation then restarted abruptly at the end of Heinrich 1. Lehman and Keigwin (1992) developed this by showing that a change in sea surface temperatures controlled surface air temperatures. Their model showed changes in temperature spreading a far as central Europe.

An ice sheet response to Heinrich Events is shown across the North Atlantic, with literature focused largely on Fennoscandinavia and the Irish Sea Basin. An advance of the Fennoscandinavian Ice Sheet, the Tampen Readvance, is widely believed to correspond with the Dimlington Stadial (Serjup *et al.* 2000). This readvance has been correlated to the Greenland ice record, showing that H1 is lagging 500-1000 years behind major ice recession suggesting that the Fennoscandinavian Ice did contribute to the Heinrich events. Nygard *et al.* (2004) suggested evidence from the Måløy Plateau indicated an ice sheet readvance shortly after 18 085 – 18 556 cal. Years BP, in the form of the Bremanger Moraine. Climatic forcing was suggested as the cause of this event due to the large scale of the readvance and was, hence, correlated with H1. Evidence was also presented indicating that H1 related conditions can be seen in the Faroe-Shetland channel and on the Vøring Plateau, as forams show very light $\delta^{18}O$ values between 13 079 – 15 975 cal. years BP. The Sharpres Event has also been correlated with H1 (Bakke *et al.* 2005) as occurring between 15 975 - 14 293 cal. years BP. This study also suggested that there was a decrease in winter precipitation during H1 shown by the record in Strupskardet indicating that the atmospheric circulation responded to the lower sea surface temperatures during H1 (Bakke at al 2005). Nygard *et al.* (2004) suggested that a synchronous response of the Fennoscandinavian and British Ice Sheets may not occur due to the size and potential response time of the ice sheets. However, the dates presented in this paper broadly correspond with those available for the British Isles as will be discussed below.

Lehman and Keigwin (1992) suggested that the oceanic signal should be particularly pronounced in Great Britain due to its location on the leading edge of the North Atlantic. Therefore, Great Britain was considered to be in a good location to detect changes in temperature resulting from Heinrich Events. Lowe *et al.* (1994) suggested that the British climate records could be explained by external changes in the Gulf Stream. The reduction in NADW directly affected the strength of this ocean current. Hence, as Heinrich events caused changes in the NADW, the Gulf Stream was also weakened. As a result British climate was cooled implying precipitation would fall as snow and ice sheets could potentially readvance. The changes in atmospheric and sea surface temperatures along with location indicated that a British Ice Sheet response was highly likely, and has been investigated with renewed interest in the past ten years.

McCabe and Clark (1998) constrained the age of an ice sheet readvance at approximately 15 150 – 14 293 cal. years BP, dates which have been supported by cosmogenic dates (Bowen *et al.* 2002). McCabe *et al.* (1998) then demonstrated a readvance of the Irish Sea Ice Stream (ISIS) as a result of H1 using a number of lines of evidence. Sedimentological evidence, along with radiocarbon dates, were published from the east coast of Northern Ireland that correlated an ice sheet advance, the Killard Stadial, to H1. A geomorphological link for this event between Northern Ireland, the Isle of Man and Cumbria was suggested in the form of an end moraine, between Killard Point, the Bride moraine and the St Bees moraine. Along with the primary evidence taken from Killard Point, McCabe *et al.* (1998) also reviewed previous literature from around Great Britain, mainly Scotland. This was used to draw up a map of ice readvance limits correlated to H1 (Figure 1). This paper marked the beginning of the increasing amount of work within the British Isles to correlate ice readvances from the late Devensian to H1, especially within the ISB. Merritt and Auton (2000) looked at sediments from boreholes in Cumbria suggesting that while there was evidence for a readvance around the time of H1 it was a previous advance, the Gosforth Oscillation, which had formed the large moraines in the area. Thomas *et al.* (2004), and then Roberts *et al.* (2006), also supported the H1 model with sedimentary evidence and radiocarbon dates from Jurby Head on the Isle of Man correlating with the event. Work has also continued in Northern Ireland strengthening the argument for an Irish Sea Basin readvance in response to H1 with dates from Ballycrampsy

(McCabe and Clark 2003), Cranfield Point (McCabe and Clark 2003) and Dundalk Bay (McCabe *et al.* 2007b).

While the focus of recent studies has been on the ISB there is also research that suggests an ice readvance on the West coast of Scotland that was also synchronous with H1. Radiocarbon evidence from the morainal bank at St Kilda, Outer Hebrides, indicated that ice began to retreat from the area between 18 545 – 18 772 cal. years BP and 14 566 – 13 673 cal. years BP following a minor readvance (Peacock *et al.* 1992). This is broadly correlative with a H1 signal as the only ice advance onto the continental shelf after 16 000 cal. years BP was the Killard Point Stadial (McCabe *et al.* 1998). The relation of the Wester Ross Readvance to H1 was discussed in Everest *et al.* (2006). It was suggested that if the oldest cosmogenic ^{10}Be date of 17 900 ± 2400 years (Everest *et al.* 2006) was taken into account then the Wester Ross Readvance could be correlated with the H1 advance, although this worked on the assumption that the greatest plausible age of a boulder was a better representation of the age of a moraine rather than the average boulder age. However, if the average age of 16 300 ± 1600 years (Everest *et al.* 2006) was taken then the Wester Ross Readvance may be related to non climatic controls and not H1.

While the H1 signal is mainly depicted in the ISB it has also been developed in the west of Scotland, as seen above. In addition, a signal is also seen along the east coast of Scotland. Peacock (1997) collated dates from various places in Scotland showing a potential H1 readvance, most notably at St Fergus. Peacock (2003) then presented further deglaciation dates from the Tay Estuary. Most recently McCabe *et al.* (2007) published deglacial dates from Lunan Bay and Gallowflat, on the east coast of Scotland, that they correlated to H1. These readvances have been discussed further in Section 2.4, which looks at a range of deglaciation dates from the east coast of Britain. Although this research indicates a BIIS response to H1 it must be noted that it is difficult to chronologically define an H event and thus any ice advances associated with it. Dates taken from the sediment cores in the North Atlantic do not state whether they refer to the beginning, middle or end of an event. As a result it can be difficult to convincingly correlate an ice advance with an H event.

Despite the numerous dated readvances in the ISB and Scotland no clear readvance signal has been depicted on the Yorkshire coast. Thus, the sediments and existing stratigraphies of the Yorkshire coast need to be re-evaluated.

2.3 Glacial History of Yorkshire

The glacial history of Yorkshire has been studied extensively over the past century and a half. Work has tended to focus on the LGM as evidence for earlier glaciations is fragmentary. However, the Basement Till is often considered to belong to a pre-LGM glaciation. Locations mentioned within this section are marked on Figure 2a, while Figure 2b shows the Quaternary stratigraphic logs from several sites mentioned. The line delimiting the extent of the LGM in Yorkshire on Figure 2a is based on the extent of the Skipsea Till and its equivalents (Clark *et al.* 2004) while the proposed limit of a H1 readvance is taken from McCabe *et al* (1998). However, this line is arbitrary and has not been based on any geomorphological or sedimentological data. Evidence for the timing of the LGM and other glacier margins in Yorkshire is based on a few poorly constrained dates (Evans *et al.* 2005), as is the case throughout Britain.

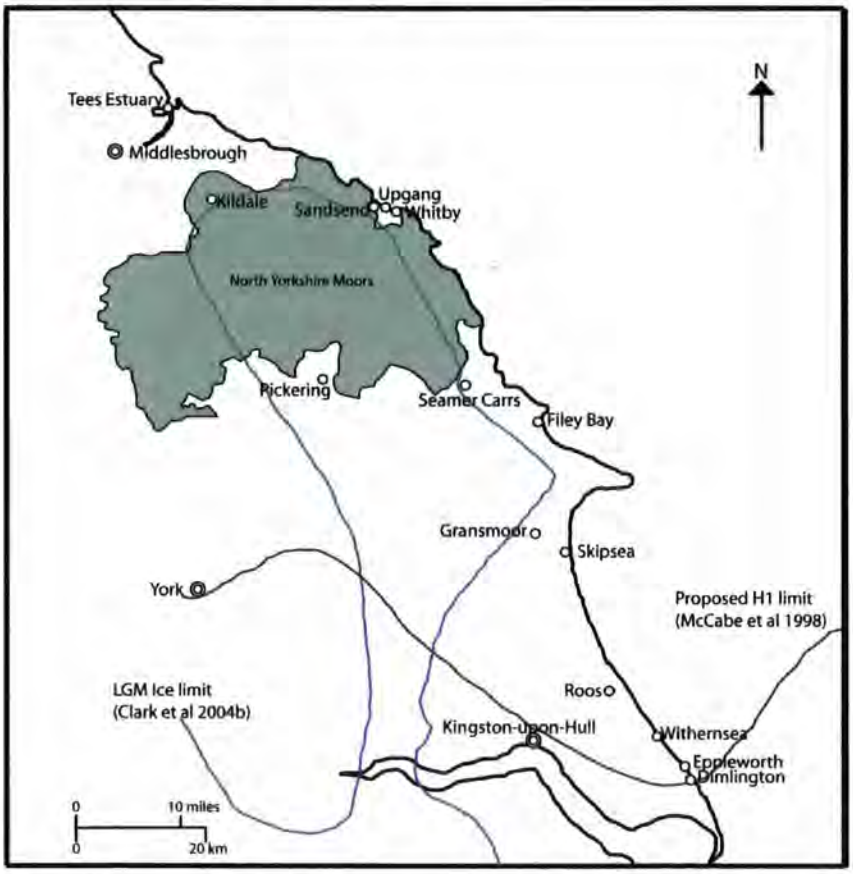


Figure 2a: Location of Upgang and other sites of importance along the Yorkshire Coast.

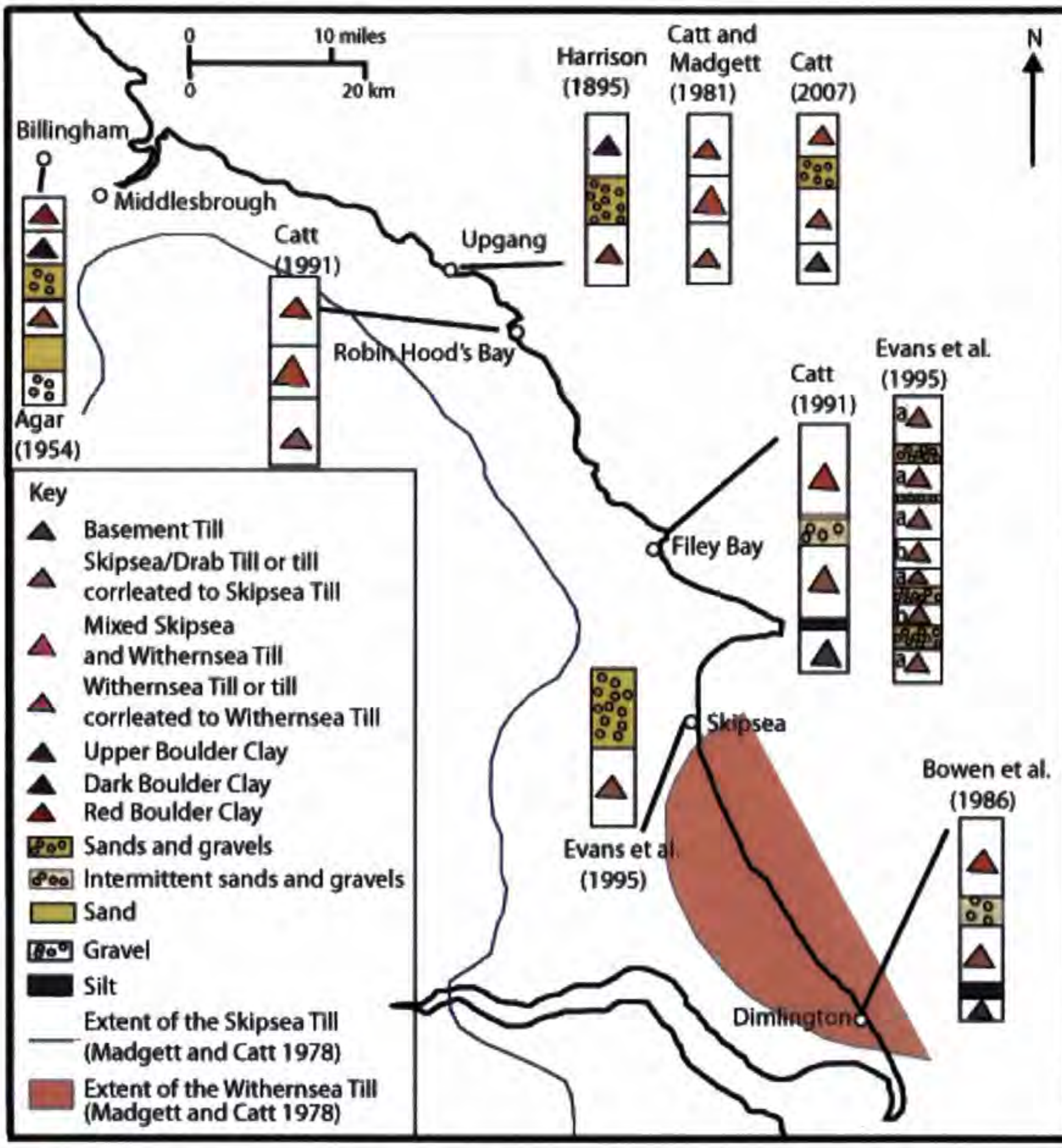


Figure 2b: Map of Yorkshire including published stratigraphic logs from sites of importance. Stratigraphic logs explained in more depth in Sections 2.3 and 2.4.

The limit of glaciation along the east coast of Britain is believed to have extended as far south as the north Norfolk coast with an ice lobe entering the Wash (Suggate and West 1959). This limit has been questioned with others stating it occurred further north, for example at Flamborough Head (Farrington and Mitchell 1951). Ice margin limits in East Yorkshire have since been reviewed and based upon a combination of sedimentology and geomorphology in East Yorkshire. Evans *et al* (2001) suggested that the limit extends onto the North Sea shelf, with ice also invading the lowlands of the East Yorkshire coast and the Vale of York. Evans *et al* (2005) discussed the evidence that was used to construct the

BRITICE database (Clark *et al* 2004a). BRITICE was a review of academic literature and maps of recorded glacial geomorphological and sedimentological features complied as a GIS database. The glacial history of three features thought to mark former ice margins, the York Moraine, Escrick Moraine and the Linton-Stutton Gravels were highlighted. The existence of several readvances was also discussed by Evans *et al* (2005), who concluded that the dating control was very weak and that several moraine systems which potentially correlated with H1 remained largely undated.

A major factor in deciphering the glacial history of Yorkshire is the extent of the Scandinavian and Scottish ice sheets interacted. Nesje and Serjup (1988) suggested that there was an ice free embayment between the two ice sheets which meant an alternative proposal would be needed to account for the deflection of the North Sea Ice Lobe south. It was also stated that while the extent of the Scottish Ice Sheet can be fairly well determined the limits of the Scandinavian Ice Sheet cannot be firmly established. Ehlers (1990) supported this model and further suggested that there was a considerable difference between the various models of ice sheet behaviour (Boulton *et al*1985; Long *et al* 1988) and the available field data. However, Ehlers and Wingfield (1991) stated that the North Sea Basin held evidence for a more extensive glaciation suggesting that the sedimentary record does not expose the full extent of the glaciation. As a result the British and Fennoscandinavian Ice Sheets could have met, thereby explaining the deflection of Scottish Ice southwards to form the North Sea Lobe (Ehlers and Wingfield 1991).

Serjup *et al.* (1994) report a radiocarbon date suggesting the coastal areas of east Scotland were ice free by 16 682 – 16 057 cal. years BP and question the possibility of a readvance in the Western sector of the North Sea correlated to the Dimlington advance. Although no evidence for such an event was found in the North Sea cores used in this paper. Carr (1999) compared the Bolders Bank Formation of the North Sea with tills in eastern England and showed that the ice which covered the southern North Sea was British in origin. He suggested that the Bolders Bank Formation is an extension of the Skipsea Till found on Holderness. Carr *et al* (2006) re-examined the extent, timing and dynamics of the glaciation of the North Sea Basin and proposed that there were three major glacial events in the region with the British and Scandinavian ice masses coalescing on at least two occasions. It is during the first two glacial events that the two ice sheets were thought to have coalesced. The final episode, the Bolders Bank episode, represented a reduced expansion of Scandinavian ice and the advance of the North Sea Lobe into the southern sector of the North Sea. Glaciological models (Boulton and Hagdorn 2007) have been created that support the existence of a fast flowing ice lobe along the east coast of Britain without the influence of Scandinavian ice, although it is glaciologically more plausible to have the North Sea Lobe buttressed by the Fennoscandinavian Ice Sheet.

The dynamics of the North Sea Lobe has been discussed with specific reference to locations on the north east coast of England. Smith (1981) considered the onshore flow of the North Sea lobe in the Sunderland District. He suggested that the thickening of ice in the Irish Sea Basin forced ice eastwards across the country and into the North Sea, where it was forced southwards due to the presence of the Scandinavian ice. Catt (1991b) looked at the movements of the lobe further south in the Tees Estuary. In this area the ice moved southwards, although a tendency for ice to move south-west in low lying coastal areas was attributed to the influence of Scandinavian Ice in the North Sea. It was however, recognised by Catt (1991a) that there is not sufficient evidence to prove that Scandinavian ice reached the southern sector of the North Sea Basin. Catt also rejects the proposal that ice from the west of Britain was deflected south by Scandinavian ice and instead suggests that the ice originated from north east Scotland and the Firth of Forth.

Work has been carried out on the Holderness coast for over a century. Lamplugh (1882) first described the tills found on this coast specifically at Bridlington. He reported a lower Basement Till at the base of the section with an overlying purple "boulder clay" and a gravel layer between the two that showed evidence of glaciotectonism. The stratigraphy of the cliff has been debated numerous times since.

The nomenclature was reconsidered and simplified by Madgett and Catt (1978) and this has become the accepted stratigraphy of the Holderness Coast. The Basement Till was considered to represent a pre-LGM glaciation due to the amino-acid ratio dates from Speeton Shell Bed in Filey Bay. The position of the shell bed under the Basement till suggested the till was deposited during Oxygen Isotope Stage (OIS) 6, the Wolstonian, as the shell bed dates to OIS 7 (Wilson 1991). The presence of the Hessle Till was dismissed as a postglacial weathering profile, while the upper till was renamed the Withernsea Till and the lower till the Skipsea Till. Both tills were dated to have been deposited after 21 000 cal. years BP in Holderness. Due to the absence of a weathering profile the two tills were considered to have been deposited by a multilayer ice sheet (Catt and Penny 1966). This is a composite ice sheet made up of ice flows superimposed upon one another flowing from different sources (Catt 1991b). The Skipsea Till is described as occupying the whole area of Holderness, while the Withernsea Till is less widespread (Figure 1b). However, a conflict appears to exist regarding the provenance of the Skipsea and Withernsea Tills. Catt and Penny (1966), Madgett and Catt (1978) and Catt (2007) stated that the Skipsea Till was deposited by ice from the Cheviots and Scotland with the Withernsea Till deposited by ice from the Lake District and Pennines. Conversely, Bisat (1939), Radge (1939) and Foster (1987) suggested that the lower, Skipsea Till was deposited by ice arriving in the area through the Stainmore Gap from the Lake District and Pennines. The Withernsea Till was believed to have been sourced from the Cheviots and southern Scotland. These ice flow directions are shown in Figure 3.

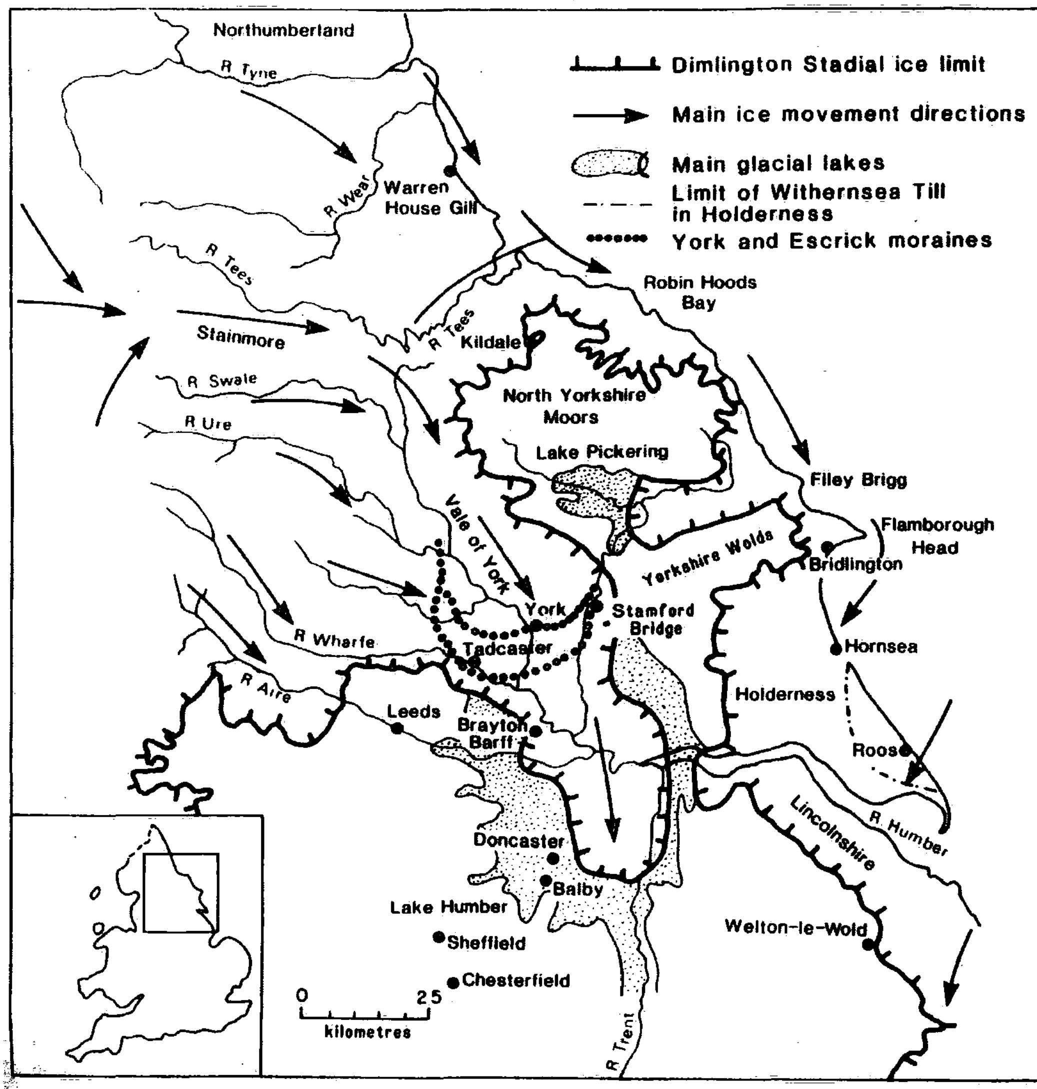


Figure 3: Map of ice flow directions and the extent of LGM ice in Yorkshire (Catt 1991b).

Foster (1987) re-examined the glacial sequence of the Dimlington Stadial in Holderness, and indicated that the previous model of a complexly stratified ice mass was inadequate. Instead it was suggested that the Skipsea till was deposited by ice moving through the Stainmore and Tees Gap from the Pennines and the Lake District. This ice was then thought to have been displaced along the coast by ice depositing the Withernsea till from southern Scotland and the Cheviots.

The chronology of the tills at Dimlington was reviewed by Eyles *et al.* (1994) who suggested the Basement Till was Late Devensian in age based on amino-acid dates, and not Wolstonian (OIS 6). The three tills at Holderness were thus classed as deformation tills deposited by ice surging onshore from the North Sea during the LGM (Eyles *et al.* 1994), with ice advances on shore marked by a series of moraine ridges.

Evans *et al.* (1995) examined two depositional models for the tills of the Holderness coast. The first one suggested that lodgement or deformation till was deposited followed by a meltout till which was then overridden. The second model, which was adopted due to the field evidence, suggested that the till was deposited by vertical accretion with associated cavity or channel fills beneath active ice. In adopting this model, Evans *et al.* (1995) questioned the validity of the previously published stratigraphies. The sands, gravels and laminated muds topping the sequence at Skipsea and Barmston were thought represent proglacial fluvial or lacustrine sediments deposited on a sandur plain during deglaciation.

While the type site of Dimlington has been studied on numerous occasions, some authors have looked at other bays on the East Yorkshire Coast to the north of Holderness. Lamplugh (1879) compared the tills north of Flamborough Head with those to the south at Holderness. The Basement and Purple Tills were seen to dominate at Holderness, while the Purple and Brown dominate in the north. Both areas were said to be capped by the Hessle till.

Catt (1991b) gave descriptions of the sediment sections along the Yorkshire coast, at Filey Bay and Robin Hoods Bay. The Basement Till there shows evidence that it was pushed into its current form by later overriding ice. He also reaffirmed the idea that the Skipsea and Withernsea Tills had been deposited by a composite ice sheet with several tributaries (Catt 1991b).

Work on the Whitby area showed a triple division of the drift that made up the cliff section between Whitby and Sandsend (Harrison 1895). These three layers were described in detail by Harrison (1895) and were shown to continue both north and south of Whitby. They were classified as Lower Boulder Clay, Middle Sands and Gravels (although this layer was not considered to be continuous) and Upper Boulder Clay, which was lighter in colour and softer than the lower till. Harrison (1895) also considered the provenance of the till proposing that none of the ice was sourced locally due to a 'snow shadow' created by the Welsh Mountains and the Pennines. A 98 metre raft of Liassic shale was found at the base of the sedimentary sequence at Upgang Beach (Hemingway and Riddler 1980). The contact of this raft with the red-brown diamict above it is described as sharp and remarkably horizontal, while the lower boundary is also sharply defined. It was hypothesised that the ice lobe became restricted as it flowed south, favouring the shearing of blocks from north face escarpments, the likely source for the shale raft (Hemmingway and Riddler 1980). It was proposed that this raft consists of shale from the Cleveland Ironstone Formation from the north, which has been driven on land and pushed upwards to its position within the Upgang sequence. Other work looking at Upgang Beach, and other sites north of Holderness, was carried out by Bisat as reviewed in Catt and Madgett (1981). Two tills were recognised at these sites often separated by sands and gravels. These north Yorkshire tills were compared to those in Holderness, with the same erratic suite appearing to be present. At Upgang an intermediate bed around eight metres thick was found between the Drab and Purple clays. This bed was considered to be a mixed till consisting of both the Drab and Purple clays. Bisat did not comment on its origin, but Catt and Madgett (1981) concluded that it could be the result of mixing during deposition from the composite ice sheet. Slightly further north along the cliffs at Sandsend another clay layer was discovered above the Purple till containing small British clasts (Catt and Madgett 1981) although the origin of this layer was not discussed further.

The glacial history of North Cleveland was later examined in Radge (1939). In this paper it was argued that the ice that invaded this area was the Cheviot Ice Stream from the northeast. Radge (1939) also discussed the deglacial history of the area with, large quantities of sands and gravel, along with a few paleoeontological finds, suggesting that large amounts of water were present upon deglaciation, with the inland ice retreating before that in coastal areas (Radge 1939).

The glacial history of Middlesbrough and the Tees Estuary was examined by Agar (1954). The whole area was described as being covered by glacial deposits. A typical section between Middlesbrough and Billingham consisted of gravel, sand, Drab boulder clay, sand and gravels, dark boulder clay, topped by Red boulder clay. Agar (1954) suggested that there was not sufficient evidence to correlate these clays with those further south. Although it was thought that the Red and Drab boulder clays could be associated with the Purple and Drab clays at Holderness and Upgang as described by Bisat (1939). This correlation along the cliffs was again proposed by Catt and Penny (1966) in which it was suggested that the Drab-Purple-Hessle series could be traced northwards toward the edge of the North Yorkshire Moors.

Also of importance in the Tees estuary is the presence of a proglacial lake during the LGM (Plater *et al.* 2000). The existence of Lake Tees implies that Lake District ice that flowed through the Stainmore Gap did not always (or indeed may never have) extend to the east coast and/or have been coalescent with North Sea Lobe ice. For an extended period during deglaciation, ice flowing along the coast from the north blocked drainage from the Lake District ice impounding a lake in the Tees estuary (Plater *et al.* 2000).

2.4 Deglacial Chronology of the East Coast of Britain

Attempts to reconstruct the deglacial chronology of Great Britain, and specifically the east coast, have used many techniques. These include radiocarbon and thermoluminescence dating, as well as correlating vegetation records with the Greenland Ice cores. Many of these deglacial dates support the idea that a glacial readvance occurred at the end of the LGM (Table 1).

Table 1: Published dates marking the deglacial chronology of the east coast of Britain.

Site	Radiocarbon Date	Calibrated Date	Site details	Literature Source
Heinrich Event 1	13 490 – 14 500	16 117 – 17617*	Not specified	Bond *et al.* (1992)
Dimlington, Yorkshire	18 240 ± 250	20 931 – 22 281 *	Moss within the Dimlington silts.	Penny *et al.* (1969)
Dimlington, Yorkshire (TL date)	N/A	17 500 ± 1.6 x 10^3 (TL)	The loess component of a solifluction deposit overlain by till.	Wintle and Catt (1985)
Kildale, Yorkshire	16 713 ± 340	19 144 – 20 537 *	Moss within a thick shell marl. (date area was ice free)	Jones (1977)
The Bog, Roos, Yorkshire	13 045 ± 270	14 360 – 12 628*	From a kettle hole within Withernsea till.	Beckett *et al.* (1981)
St Fergus, east Scotland	14 915 ± 210	17 386 – 18 727 *	St Fergus Silts, towards top of deposit.	Hall and Jarvis (1989)
Troup Head, Scotland	14 000	15 150 – 14 293*		Merritt *et al.* (1995)
Lunan Bay, east Scotland	17 065 ± 50 17 720 ± 50	20 250 ± 40 21 130 ± 90	From raised marine mud overlain by ice contact gravel.	McCabe *et al.* (2007)
Gallowflat, east Scotland	13 655 ± 45 13 675 ± 40	16 355 ± 130 16 385 ± 125	From laminated muds 1m above glacial diamict	McCabe *et al.* (2007)

* Calibrated from published 14C date using Calib Rev 5.0.1 (two sigma ranges)

A key site in the deglacial chronology of the east coast is Dimlington on the Holderness coast of Yorkshire. Two radiocarbon dates of 18 500 ± 250 ^{14}C years BP (20 559 – 19 258 cal. years BP) and 18 240 ± 250 ^{14}C years BP (20 931 - 22 281 cal. years BP) were derived from moss within the Dimlington silts which underlie tills at this site (Penny *et al.* 1969). Originally interpreted as an interglacial deposit these silts consist of laminated silts and sands at the base of the cliff. However, the analysis of coleoptera from this lithofacies suggested that the silts were deposited during the early Upper Pleniglacial (c. 29 000-15 000 ^{14}C years BP), the period of maximum glaciation on continental Europe, and not an interstadial (Penny *et al.* 1969). The moss was thought to have grown in a cold, potentially meltwater environment during this time. While Penny *et al.* (1969) did not discuss the date in relation to the glacial chronology of the area, it led Rose (1985) to propose the Dimlington Stadial as a climatostratigraphic name for the LGM in Britain. Thought to represent the main glacial episode of the Devensian between 19 000 and 23 000 cal. years BP, this stadial was considered to represent a deterioration in climatic conditions and the advance of the ice sheet. Dimlington was named as the type site as the Skipsea and Withernsea tills directly overlie the dated Dimlington silts. The date indicated that the tills overlying the silts at Dimlington were both deposited post 21 000 cal. years BP as a result of a deterioration and then amelioration of climate with the Skipsea till representing the LGM till in Yorkshire. Potentially, this was accompanied by minor oscillations in the ice front. These dates for the LGM on the east coast broadly correspond to recent radiocarbon dates from the ISB which suggest that the LGM can be dated to c.24 000 cal. years BP (O'Cofaigh and Evans 2007), but could suggest a slight diachroneity between west and east coast sectors of the BIIS. O'Cofaigh and Evans (2007) suggested that the Dimlington Stadial represents a readvance of an unstable North Sea Lobe reinforcing the view that the British and Irish Ice Sheet was highly sensitive during the LGM to climate, sea level and internal dynamics.

Dating of the tills along the Holderness coast has also been carried out using thermoluminescence dating (TL) on loess within a solifluction deposit below the Skipsea till at Eppleworth (Wintle and Catt 1985). The TL date obtained from the silt fraction of this deposit was 17 500 ± 1 600 years BP. However, the high chalk content present would have diluted the radioactivity. Therefore, the sample was also dated after treatment with HCL resulting in an age of 16 600 ± 1 700 years BP (Wintle and Catt 1985).

McCabe *et al.* (1998) examined these published dates and proposed that they could be correlated with H1. It was suggested that the original dates on the Dimlington mosses had been called into question by thermoluminescence dates on the solifluction deposit underlying till. As the dates were taken from below the Skipsea Till at Eppleworth, non glacial conditions were thought to have existed around 17 500 cal. years BP, and thus, the Dimlington dates were too old (McCabe *et al.* 1998). Therefore, the Skipsea and Withernsea tills of Holderness were considered to have been deposited during the Killard Point Stadial of North Ireland. Peacock (1997) also agreed with this view as the calendar thermoluminescence dates correspond very approximately 15 to 14 ^{14}C ka. However, if both the Skipsea and Withernsea Tills were seen to be H1 readvance tills then there does not appear to be an 'LGM' till in Yorkshire. While a large proportion of the literature refers to the Skipsea and Withernsea Tills as the two LGM tills in Yorkshire, Eyles *et al.* (1994) offered an alternative proposition. It is claimed that amino-acid dates taken from the Basement Till at Dimlington date the till to be LGM in age. Therefore, if the Skipsea and Withernsea Tills are taken to be H1 tills then the Basement Till would be the sole LGM till.

Another radiocarbon date from the LGM is from a bone fragment found in the "older littoral sand and gravels" near Brantingham, East Yorkshire (Gaunt 1974). This date of 21 834 ^{14}C years BP represented the date for the initiation of glacial Lake Humber or the date of the lakes high stand depending on whether it comes from the base of the sands and gravels or from within them. Work was carried out in the Vale of York by Gaunt (1976) which was used to correlate the evidence found here with H1 by McCabe *et al.* (1998). Evidence from the area was used to conclude that ice advanced into the south eastern part of the Vale of York during the same time as the high stand of Lake Humber. However, the dates taken from Dimlington of 21 000 to 22 000 cal. years BP suggested that Lake Humber had not been initiated by this time as the North Sea Ice Lobe had not extended far enough south. This contradicted the work of Gaunt (1974).

Jones (1977) presented a radiocarbon date of 16 713 ± 340 ^{14}C years BP (19 144 – 20 537 cal. Years BP) from Kildale, north east Yorkshire. This was obtained from the lowest deposit of moss fragments above LGM till, suggesting the main deglaciation in the area occurred post 19 000 cal. years BP. The lower sections of this site had very few macrofossils or pollen data. From this it was inferred that climatic conditions were severe after the ice had retreated, with the lake that formed post deglaciation being permanently frozen. Organic layers found above this were considered to have been deposited rapidly although no dates were available. Keen *et al.* (1984) suggested that the date of c. 21 000 cal. years had been affected by hard water error making it erroneously old. A vegetational history of the area was proposed showing the progression from Late Devensian vegetation to the Holocene. The succession at the lower end of the section is discussed in more detail in Innes (2002). A clay, shell marl, clay, shell marl, sedge peat succession is described. It is from the lowest shell marl that the date of 19 144 – 20 537 cal. Year BP was obtained. It is therefore possible that ice readvanced toward this area after 19 144-20 537 cal. years BP depositing the second clay unit, before an improvement in climatic conditions resulting in the second unit of shell marl. However, it was again suggested within Innes (2002) that the date was too old owing to hard water error. Despite the suggestion that this date was erroneously old, the dates are not inconsistent with other radiocarbon dates from similar stratigraphic positions in other parts of Britain (Rose 1985).

Another deglacial date from Yorkshire comes from Roos Bog near Dimlington. This showed the area to be ice free by 14 360 – 12 628 cal. Years BP (Beckett 1981). In conjunction with the Dimlington dates this gives a maximum of approximately 5000 years for the deposition of both the Skipsea and Withernsea tills.

An alternative method used to date the glaciations of Yorkshire and the east coast has been 'wiggle matching' local climate reconstructions with the dated Greenland ice cores. Mayle *et al.* (1999) presented data from different areas in Britain in comparison with the GRIP ice core record between 15 000-11 500 cal. years BP. These records from South Wales, Yorkshire, Southern Scotland and Northern Scotland showed a degree of compatibility with the GRIP ice core record. Climate was shown to have warmed after 15 000 years BP, before cooling commencing at 14 500 years BP. The climate was then unstable before climatic cooling at the LLS. These records showed a good correlation with the Greenland ice core record, implying that British climate can be inferred from them in the absence of dated climatic records from sites in Britain. Alternatively this correlation can be used to 'date' British climate records by the process of 'wiggle matching' them to the Greenland records. A time lag could be present between the climate and vegetation records as suggested by Walker *et al.* (1993) in Gransmoor, East Yorkshire. This vegetation-climate lag at the beginning of an interstadial was thought to be several hundred years as plants colonised a previously barren area (Walker *et al.* 1993). Therefore, it may not be possible to directly correlate the vegetation and ice sheet records in Yorkshire.

A correlation between the east coast geological record and H1 is also found within the Tees Estuary. Plater *et al.* (2000) drew a comparison between a varve sequence and the GISP ice core record. This varve record was thought to have been deposited in proglacial Lake Tees, which was impounded by the North Yorkshire Moors to the south and ice masses to the North and North East. A good correlation between this core and the GISP record was shown, indicating that sediment supply to the lake was controlled by changes in climate, and thereby creating a climate proxy. A deterioration in temperature was seen around the same time as H1. In addition, this core was luminescence dated to 18 365 ± 10 015 years BP (Plater *et al.* 2000). Nevertheless this date has a very large error term attached to it, and does not match the date proposed by the correlation with the GISP record. This uncertainty is attributed to water content and water content history reconstructions (Plater *et al.* 2000).

However, the thickness of the unit suggested that the record may be influenced by the proximity to the lakes shoreline and local deltaic deposits. Therefore, only the period of 18 000 -16 000 years BP was chosen for correlation as the varve sequence from this time must have been deposited during a period of ice impoundment prior to the deglaciation of the area (Plater *et al.* 2000). Thus, the influence of the lake margin and any deltaic deposits was minimised.

Outside Yorkshire and the Tees Estuary other sites can be considered in order to establish the deglacial chronology of the east coast. The maximum limit of the last glacial ice sheet was believed to be just onshore of the North Norfolk coast (Brand *et al.* 2002). Therefore any chronology along this limit will provide the date of maximum glaciation or initial deglaciation. This limit has been mapped by looking at the extent of the Holkam Till (Bowen *et al.* 1999), previously named the Hunstanton Till (Sudgate and West 1959). Brand *et al.* (2002) did not present directly date any sediments, however, it was suggested that the till found in North Norfolk had an affinity with the glacial deposits of Holderness, and hence, could be dated to post c 20 559 – 19 258 cal. years BP. The Holkam till overlays a raised beach that has been dated to oxygen isotope stage 5e, confirming the interpretation of the till as Devensian (Gale *et al.* 1988). Evidence from north Norfolk was also used to support the model of a North Sea Ice Lobe flowing along the east coast of England from southern Scotland (Pawley *et al.* 2006). Therefore, if two ice lobes, one from the Lake District and one from Scotland, flowed along the coast of Yorkshire only one, from Scotland, reached Norfolk.

Deglacial dates have also been presented from the east coast of Scotland which suggested a readvance potentially correlating with H1. Radiocarbon dates suggested that a readvance of ice at Troup Head, east of Macduff occurred prior to 15 150 – 14 293 cal. years BP (Merritt *et al.* 1995). Other evidence of readvances of the Moray Firth Ice Stream was also presented by Merritt *et al.* (1995). However, they suggested that changes in relative sea level controlled ice stream behaviour and not climate.

Peacock (1997) furthered this by suggesting that terrestrial ice reached a maximum extent in the North Sea between 17 000 and 16 000 cal. years BP. A large moraine at St Fergus was dated to 17 386 – 18 727 cal. years BP (Hall and Jarvis 1989). Peacock (1997) suggested that this evidence along with ice sheet movement directions at Fraserburgh showed an ice sheet readvance into the North Sea, contemporaneous with H1. He also suggested that this conclusion was to be viewed with caution as at the time of publishing no dates suggesting a H1 readvance had been published in Ireland. It was recognised that there was no direct evidence for a major late Devensian ice readvance at Dimlington which would be synchronous with H1.

Further deglacial dates from eastern Scotland where taken from the Tay Estuary (Peacock 2003). These dates suggested that deglaciation of the area began around 16 000 -16 500 cal. years BP resulting in two hypothesis. Firstly, if the ice started retreating at 15 150 – 14 293 cal. years BP then the Tay glacier was retreating during H1. As a result only a weak H1 signal can be proposed as may be the case around the entire North Sea. Secondly, if the 16 000 cal. years BP date is taken to represent deglaciation then there was time for the ice to retreat into the Scottish highlands before readvancing. This could correlate with H1 and the Perth readvance limits proposed by Sissons and Smith (1965). Peacock (2003) also emphasised the need to exercise caution when assigning readvances to the H1 event as a result of the varying dates available around the British Isles that had been correlated with the event.

This proposal has been supported by a revised model for the deglaciation of the east coast of Scotland (McCabe *et al.* 2007). Radiocarbon dates are used to support previously published sedimentary data indicating two glacial readvances before the beginning of the Loch Lomond Stadial. The first readvance, the Lunan Bay readvance was dated to after 20 250 cal. years BP indicating initial deglaciation had begun prior to this date. The second readvance, the Perth readvance, was dated using two radiocarbon dates from Gallowflat of 16 355 ± 130 cal. years BP and 16 385 ± 125 cal. years BP. This readvance was thus correlated with the Killard Point Stadial proposed in North Ireland (McCabe *et al.* 2007a) and supported by the dates from the St Fergus silts and those from Peacock (2003).

Along with the reconstruction of deglacial chronologies on land, attempts at dating the offshore record have also occurred. Serjup *et al.* (1994) dated the maximum extent of Devensian ice in the North Sea to the period between 22 000 and 29 000 ^{14}C BP, with the central sector ice free around 22 000 ^{14}C BP. A possible readvance of ice in the area occurred around 21 000 cal. years BP, correlating with the Dimlington Stadial of Britain and the Tampen Readvance of Norway. Difficulties exist in trying to constrain the age of the last glacial period in the North Sea. Radiocarbon dates taken from a marine setting carry uncertainties as a consequence of varying reservoir ages. This results in difficulties in comparing terrestrial and marine events (Peacock 1995).

2.5 Field Site

The sediments of Upgang will be examined to see if a readvance signal can be established and correlated with H1. Previous literature published in relation to Upgang suggests that the site has two tills separated by sands and gravels and, hence, may show a readvance (Section 2.3), although this research has resulted in a number of different stratigraphies (Figure 1b). Harrison (1895) reported a lower and upper boulder clay separated in places by a non continuous sands and gravels facies. Catt and Madgett (1981) described two tills correlating to the Skipsea and Withernsea till often separated by sand and gravel with the addition of an intermediate bed of mixed Skipsea and Withernsea Till. Most recently Catt (2007) also described Skipsea and Withernsea Till equivalents separated by sands and gravels. However, in this interpretation there was no intermediate bed and a lower grey till correlated to a pre-LGM glaciation was depicted. In addition, Upgang's location along the northern Yorkshire coast will help to provide a link between on going research in County Durham and the Tees Estuary with that carried out to the south in Holderness.

Upgang beach is located approximately 1 km north of Whitby just south of Sandsend. It is situated on the edge of the topographical highland plateau of the North Yorkshire Moors National Park (Figure 2) lying on a Jurassic bedrock or shales, clays, limestone and sandstone. Topography in the Upgang and North Yorkshire area is dictated by this bedrock and consists of dissected hills and wide moorland (Kent 1980). Upgang lies on the high irregular cliffs indented by small bays and narrow valleys that characterise the coastline between Saltburn and Flamborough Head (Kent 1980).

The cliff section is approximately 750 metres long and 30 meters high and is composed of diamicts, sands and gravels. The cliff section is kept relatively 'clean' by erosion, up to 30 cm per year; with long shore drift southwards (English Nature 2006). However, there are areas of slumping obscuring the *in situ* sediments.

2.6 Geology of the area

The on and offshore geology for eastern England can be seen in Figure 4. The bedrock in North East Yorkshire is Jurassic and runs along the coast between Middlesbrough and Filey. This sequence consists of shales and clays with limestones, ironstones, siltstones and sandstones (Kent 1980). Shales are the most dominant lithology with limestone, ironstone and sandstone outcrops occasionally occurring in the Lower to Upper Lias part of the succession. This section of the geological sequence contains regular fossils including *Ammonites* and *gryphaea arcuata* (devils toe nail). The Middle Jurassic element of the succession is characterised by the Dogger unit, a marine unit comprising of similar lithologies to the underlying Lias sequence although locally oolitic. Above the Dogger formation the geology begins to be dominated by sandstone with some shales, the Saltwich formation, and Eller Beck formation consisting of shales and shaley sandstone.

To the north of the Jurassic sequence, around Middlesbrough, the solid geology is dominated by Triassic mudstones which to the north and east become Triassic sandstones. Following the solid geology north along the coast line between Hartlepool and South Shields is Permian limestone. Between South Shields and Amble are the coal measures followed by Carboniferous limestone to the north in Northumberland (Figure 4). A major part of the Northumberland geology is the Whin Sill complex, a series of sills and dykes of basaltic composition with the most common lithology a dark-blue-grey quartz dolerite.

Further north towards the Cheviot Hills the bedrock consists of igneous rocks. This is dominated by dark grey and pink granite. There are also outcrops of dark grey or black andesite, rhyolites consisting of abundant feldspar and biotite phenocrysts and grey or purple porphyry with white or pink feldspars.

Igneous rocks also occur to the west of the north east in the Lake District. These are dominated by various granites of varying colour and composition. Grey, dark green/purple andesites and grey fine grained rhyolites are also sound in the area. To the east of the Lake District the geology of the Pennines is dominated by large areas of Namurian grit, with regions of slate, sandstones and limestone.

Overlying the Jurassic bedrock in Yorkshire is a series of Quaternary glacial sediments. It is these sediments that have lead to the numerous debates discussed in Section 2.3.

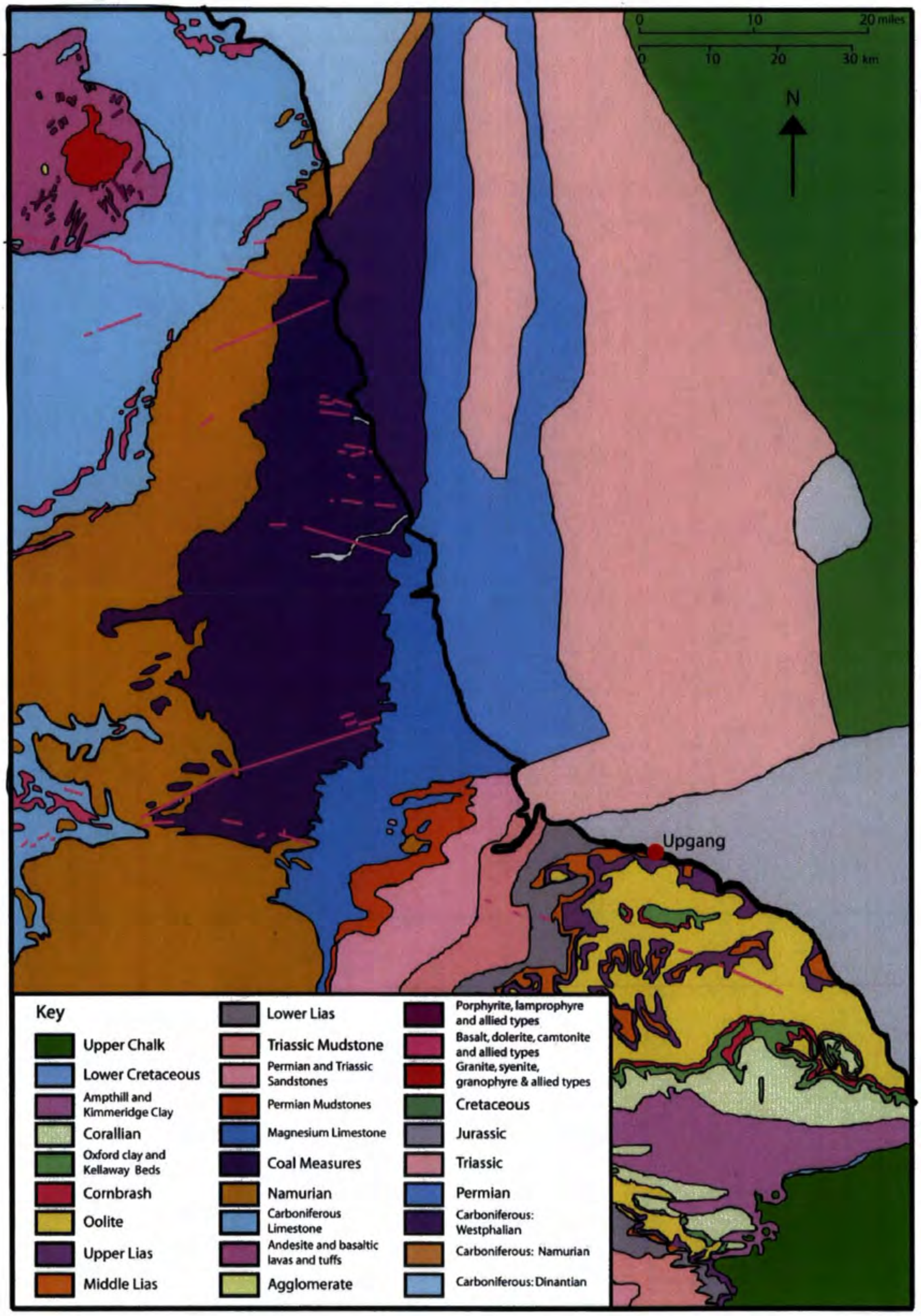


Figure 4: Geology of North East England (From BGS Solid Geology Map (North and South) and BGS Offshore Solid Geology Map (North and South).

3. Methods

3.1 Sediment Description and Lithofacies Associations

The accurate description of the sediments present at Upgang provides the framework to this investigation. It is one of the most important and useful tools in glacial reconstruction (Evans and Benn 2004). Sediment facies often display a distinct set of characteristics depending on mode of deposition. This allows the depositional history of a site to be reconstructed, although, this interpretation is carried out at a later stage. Logging is purely descriptive and objective taking note of sediment colour, grain size, depositional and deformational structures, bed thickness, contacts between lithofacies and the presence of any inclusions. This was done by taking sketches, photographs and notes. A more detailed description of this process can be found in Jones *et al.* (1999). Lithofacies codes have been used to allow rapid and effective description of the sediments in the field. A hybrid scheme has been used encompassing the code proposed by Kruger and Kjaer (1999) to describe the diamictic sediments, while the sorted material was described using the code in Eyles (1983) (Figure 5). The combination of the two codes allows sediment to be described in greater detail. Kruger and Kjaer (1999) provided a more detailed code for the description of diamicts than Eyles (1983), while the sorted sediment code was more descriptive and wide ranging in Eyles (1983). The Munsell Colour Chart was used to objectively describe the colour of sediments (Graham 1988).

For consistency, a hierarchy is used when grouping sediments together in description. The largest group that was used in this report is a Lithofacies Association (LFA). This is a mappable layer of homogenous material (Evans and Benn 2004). Each LFA can then be subdivided into facies which can be divided further into units if necessary. For example, a layer of sand and gravels would be classed as an LFA. This may then be sub divided into a banded sand facies, each band within this facies would be a unit.

Code for diamicts (from Kruger and Kjaer 1999)

D	Diamict
General appearance:	
m	Massive, homogenous
g	Graded
b/s	Banded/stratified
h	Heterogeneous
Granulometric composition of matrix:	
C	Coarse-grained, sandy-gravelly
M	Medium-grained, silty-sandy
F	Fine-grained, clayey-silty
Clast/matrix relationship:	
(c)	Clast-supported
(m_1)	Matrix-supported, clast poor
(m_2)	Matrix-supported, moderate
(m_3)	Matrix-supported, clast rich
Consistency when moist:	
1	Loose, not compacted
2	Friable, easy to excavate
3	Firm, difficult to excavate
4	Extremely firm

Code for sorted sediments (from Eyles *et al.* 1983

Gravels	
Gms	Matrix supported, massive
Gm	Clast supported, massive
Gsi	Matrix supported imbricated
Gmi	Clast supported massive
Gh	Horizontally bedded
Gt	Trough cross-bedded
Gp	Planer cross bedded
Gfu	Upward fining
Gcu	Upward coarsening
Granules	
GRcl	Massive with clay laminae
GRh	Horizontally bedded
GRm	Massive and homogenous
GRcu	Upward coarsening
GRuf	Upward fining
GRp	Cross bedded
Sands	
St	Trough cross bedded
Sp	Planer cross bedded
Scr	Climbing ripples
Ssr	Starved ripples
Sl	Horizontal and draped lamination
Sm	Massive
Suc	Upward coarsening
Suf	Upward fining
Silts and Clays	
Fl	Fine laminations
Fm	Massive

Figure 5: Lithofacies codes

A number of detailed logs were taken of both vertical sections, highlighting the changes in sediments up the sequence, and horizontally at sites of interest. A horizontal overview of 750 metres of the cliff was also taken highlighting the key sedimentary facies, allowing sample sites and the sites of detailed logs to be placed in context.

The main limitation with this method arises during interpretation. It is possible for the depositional history of a sediment to be interpreted in a number of different ways. This is the result of different processes resulting in the same sedimentary sequence, equifinality. However, if used in conjunction with other complimentary techniques a reliable reconstruction can be created.

3.2 Clast Form Analysis

Clast form analysis involves measuring and determining the roundness of individual clasts to establish the shape. This helps to establish the depositional history of the sediment. The long (a), intermediate (b) and short (c) axis of the clasts are measured using a standard tape measure (Evans and Benn 2004, Hubbard and Glasser 2005). A sample of fifty clasts needs to be taken for the data to be statistically significant within an area of one metre squared as differences can occur even in small areas as a result of changes in depositional processes. As far as possible similar lithologies should be measured to help reduce inconsistencies within the sample as different lithologies have different weathering rates and material properties. In addition similar sizes of clasts were measured (8-40 mm). Problems can occur in clast poor sediments as there may not be fifty clasts of a similar lithology within a metre square. Therefore it may be necessary to enlarge the sample area or to reduce the limitations on lithology. Once the clasts have been measured the roundness needs to be evaluated. This was done using the Powers roundness scale (1953). While other classification techniques have been suggested (Lees 1964, Olsen 1983), the Powers roundness scale is quick and easy to use in the field without involving extra measurements. It is based on a six point scale with clasts classified as very angular, angular, sub-angular, sub-rounded, rounded or well rounded. Despite the advantage of speed within the field the method can be subjective. This data is presented in a number of ways. The measurement data is presented in ternary diagrams (Sneed and Folk 1958). These plot the a, b and c axis between three forms; spheres, discs and rods (Benn 1994). The benefit of these ternary diagrams over other ways to display the data (cf. Hoffman 1994) is that they represent the continuum of a particle shape with equal weighting and relative simplicity (Benn and Ballantyne 1995). Roundness data was plotted on a histogram to allow clear visual representation of the data.

A covariance plot is also constructed using this data in order to assess the transport history of each sample. The C40, the number of clasts below the 0.4 line on the ternary diagrams, and RA, the number of angular and very angular clasts, values of each sample is calculated and then plotted onto a graph. This graph can then be compared with published data sets which differentiate between various transport pathways (cf. Benn 1992, Benn and Ballantyne 1994). Scree and supraglacially transported material will plot to the top right corner, while river deposits and fluvially transported samples will plot to the bottom left.

3.3 Clast Fabric Analysis

Clast fabric analysis was under taken in order to support clast form data in establishing deposition history. It involved measuring the dip and azimuth of individual clasts within a sample (Evans and Benn 2004, Hubbard and Glasser 2005) using a compass and clinometer. Again a sample of fifty clasts was measured to reduce random variability (Ringrose and Benn 1997) within a one metre squared area. Clasts with an a axis 1.5 times as long as the b axis should ideally be used as these present a preferred orientation. Clasts of a similar size were measured (8 – 40 mm) as it has been shown that fabric data can vary with the size of a clast (Kjaer and Kruger 1998, Carr and Rose 2003). Measurements were predominantly taken on the a-axis of the clasts; however, for comparison, measurements were taken using the a-b plane at six sample sites. A-b plane fabrics were taken as well as a-axis fabrics. Some authors believe this method to show a clearer strain signal than a-axis fabrics that can align perpendicular to flow (Evans *et al.* 2007). A-b plane measurements were focused on areas just above a boundary as this is where the strain signal will be clearest. Data is presented using rose diagrams, indicating ice flow direction and equal area stereonets which have been contoured through the step function and Gaussian weighted methods using the Rockworks program.

The usefulness of this method has been considered within the literature. Bennett *et al.* (1999) question the value of this technique as a method to distinguish between two or more tills as different till facies do not always have distinct clast fabric signatures. While it is concluded that clast fabrics alone cannot be used to differentiate between tills it is suggested that it can be useful as a relative strain indicator. Its wide use by a number of glacial researchers cannot be ignored and if used in conjunction with other methods it can prove useful in glacial reconstruction.

In addition to the presentation of the raw data on rose diagrams and stereonets, eigenvalues were calculated using the method set out in Mark (1973). This allows a large data set to be reduced to simple descriptive statistics that describe fabric strength and orientation (Benn and Ringrose 2001). Accounting for both dip and orientation these values are calculated using vector analysis to extract eigenvectors (V1, V2 and V3) and eigenvalues (S1, S2 and S3) from a 3x3 matrix (Dowdeswell and Sharp 1986). The eigenvector V1 reflects the direction of maximum clustering, while V3 reports the direction of minimum clustering. The eigenvalues represent the strength of clustering around the axis of V1, V2 and V3. These values are then plotted. Firstly as a ternary graph (Benn 1994) scaled using an isotropy index ($I = S3/S1$) and an elongation index ($E = I - (S2/S1)$) and, secondly, as an x-y graph with S1 plotted against S3 (Dowdeswell and Sharp 1986). Once constructed these graphs can be compared with those in the published literature in order to establish till type, potential degree of deformation and fabric 'shape'. However, there is no complete theory that describes the variance of fabric eigenvalues or definitive confidence regions that can be plotted onto the graphs (Benn and Ringrose 2001). Although Benn and Ringrose (2001) use a bootstrapping process to show that the variation within samples is smaller, and often a result of random sampling effects, than the variation between till units. Therefore, when used in conjunction with other techniques this method can provide a useful tool in the differentiation of till depositional histories.

3.4 Clast Lithology Analysis

Clast lithological analysis was under taken in order to establish the provenance of the tills. The lithology of the tills at Upgang can be correlated with local and regional geologies in order to establish where each till originates from. It has also been used to help differentiate between diamict layers and help establish to which, if any, diamict layer the sands and gravel sequence can be correlated to. In order to do this 300 clasts were examined from each layer, this is the minimum number needed to make the sample statistically significant (Bridgland 1986). Clasts were then washed, dried and sieved into size fraction (>32mm, 16<32mm, 8<16mm and 4<8 mm) in the labs before classification. Classification used a low powered microscope in order to identify grain sizes, shapes and composition of the rock, with rocks often broken open to expose an unweathered face. In order to help identify limestone 10% hydrochloric acid was used as it reacts with the carbonate in the rock. Identification used Gale and Hoare (1991), Evans and Benn (2004) and Stow (2005).

The major source of weakness within this technique is the identification as it can often be subjective (Evans and Benn 2004). Another limitation with this method is that the weaker clasts, for example chalk or shale, may be crushed or crumbled during transport and deposition, as well as during excavation, especially when using a geological hammer to remove clasts from a solid till, leading to a non representative sample. Great care needs to be taken in removing clasts if this is the case. However, these lithologies can also continue to break up in transportation, washing and sieving, giving an elevated number of clasts than was originally sampled.

In order to attempt to create a lithochronostratigraphy samples were also taken from Dimlington (*cf.* Lee *et al.* 2004). Ice flow pathways within Yorkshire, suggest that ice originated from either the Lake District or northern England, flowing along the coast to Dimlington. Hence, the ice depositing tills at Dimlington would have had to overrun the Upgang area. Therefore, if similar clast lithologies can be established for the two sites then the tills at Upgang can be correlated with those at Dimlington. This would suggest that the two sites are of a similar age with all correlating tills deposited post 21 000 cal. years BP, creating a relative chronostratigraphy for Upgang. However, with clast lithology studies it is often only the most durable clasts that can be transported across long distances with most clasts derived locally.

Clast lithology has also been used to help differentiate between diamict layers and help establish to which, if any, diamict layer the sands and gravel sequence can be correlated to. Data was statistically analysed using the chi squared test (Davis 2002) in order to test the potential correlation between sample sites within and between lithofacies layers. Chi-squared (χ^2) testing calculates the variance of each sample from the mean. Therefore, if a small number is obtained the samples are similar to the mean and hence each other, a large number indicates a greater variance.

$\chi^2 = (O-E)^2/E$ O = Observed value, E = Expected value

In order to tell how significant the chi-squared value is, the reduced chi-squared value has been calculated, dividing χ^2 by the number of degrees of freedom (n-1). In this instance the closer the reduced chi-squared is to 1, the stronger the correlation. Using the Upgang data set, the chi-squared and reduced chi-squared values have been calculated using the whole sample, lithologies making up more than 20% of the sample (limestone and shale) and lithologies making up on average more than 10% of the sample (limestone, shale and sandstone) in order to compare the values obtained. Chi-squared testing does not work well when the expected frequencies are too low, as can occur with data sets with a large number of zeros in it, characteristic of lithological counts. It was therefore necessary to vary the number of lithologies used in calculations in order to minimise the percentage of lithologies with an expected value below five. It has been used on samples within and between each lithological unit in order to test for correlation. In addition, tests have been carried out on both the percentage and absolute values of each lithology. The Dimlington samples will also be 'correlated' with the Upgang tills using this method.

3.5 Geochemical Analysis

Another technique used to reconstruct the glacial history of Upgang was geochemical analysis. This involved collecting samples of diamict from the site and analysing the abundance of elements found within it using an ICP-MS (Inductively Coupled Plasma - Mass Spectrometer). It was then used to differentiate between diamict layers or to correlate with other sites for provenance testing, however, few sites have been geochemically analysed in Great Britain. Most examples of the use of geochemical analysis are from Canada (e.g. Broster 1985, Shilts 1993), where it is used to spatially differentiate between tills. This project used geochemistry from a vertical section to compare tills at Upgang and the Tees Estuary. The Tees Estuary could prove to be a significant area in reconstructing the glacial history of Yorkshire as the ice is believed to have flowed through this area from either the north or the west on its way down the coast. There is also a climatic record from the Tees Estuary (Plater *et al.* 2000) that could be linked to Upgang if the two sites can be geochemically correlated.

Before the sample could be analysed it needed to be prepared in the laboratory, using a standardised method. Firstly, the sample was reduced in size, to between 1 and 20g (Gale and Hoare 1991), taking care to sample only the matrix (grains <2mm) and to avoid contamination. Samples were then placed in a minus 80°C freezer overnight before being placed in the freeze dryer for twenty-four hours. These samples were then placed in the ball mill to grind the particles into uniform size. Care needed to be taken at this stage to avoid cross contamination between samples, so after each sample the milling equipment was carefully cleaned. Samples were then disaggregated using sodium hexametaphosphate before microwave digestion to remove any organics. Finally, the sample was placed in the ICP-MS machine which determined the relative abundance of isotopes due to the differences in atomic mass (Blatt and Tracy 1996). Two runs were carried out one for high abundance and one for low abundance elements (Table 2). This was then converted into a list of elements, with isotopes of each element combined, and their quantities found within the sediments. Major elements can be traced to within 200-300 ppm while the minor elements can be assessed to an accuracy of less than 5 ppm (Evans and Benn 2004). A major concern with interpreting geochemical data is establishing which part of the signal is due to provenance changes and which part is a result of changes in geochemistry as a result of post depositional weathering (Shilts and Kettles 1990).

Table 2: Table of elements examined in geochemical analysis.

High Abundance Run	Low Abundance Run
Aluminium (Al)	Silver (Ag)
Calcium (Ca)	Arsenic (As)
Iron (Fe)	Boron (B)
Potassium (K)	Barium (Ba)
Magnesium (Mg)	Beryllium (Be)
Sodium (Na)	Bismuth (Bi)
Titanium (Ti)	Cadmium (Cd)
Phosphorous (P)	Cobalt (Co)
Silicon (Si)	Copper (Cu)
	Chromium (Cr)
	Lithium (Li)
	Manganese (Mn)
	Molybdenum (Mo)
	Nickel (Ni)
	Lead (Pb)
	Antimony (Sb)
	Selenium (Se)
	Strontium (Sr)
	Thallium (Tl)
	Vanadium (V)
	Zinc (Zn)
	Cerium (Ce)
	Gallium (Ga)
	Niobium (Nb)
	Rubidium (Rb)
	Tin (Sn)
	Thorium (Th)
	Uranium (U)
	Yttrium (Y)
	Zirconium (Zr)

Data has been statistically analysed using chi-squared testing as described in Section 3.4 and cluster analysis in order to establish any differences or correlations between samples. In using cluster analysis groups were created 'clustering' samples with similar suites of elements using the statistical program Stata, which were then plotted as a cluster dendrogram which graphically displays the results. Clustering used Q type analysis as this compares samples (sites) rather than variables (elements). Complete linkage clustering is used as the longest distance between groups is used to form clusters. Unlike single linkage clustering (the shortest distance between clusters) this reduces the likelihood than an outlier will cause two groups to cluster even though overall they are distinctly different. Therefore, many smaller clusters tend to form and they are often more informative. The y axis of the dendrogram represents the Euclidean, or ordinary, distance between clusters rather than a relative distance. The closer the samples cluster to zero the more similar they are.

3.6 Geomorphologic Mapping

Geomorphological mapping highlights the spatial distribution of landforms, accounts for their genesis and can provide the relative ages of their relief (Hubbard and Glasser 2005). Therefore the creation of a geomorphological map will provide a greater context for the sedimentological part of this study.

The geomorphological map constructed for this project was created using remote sensing techniques. A Digital Elevation Model (DEM) was created of the North East and Yorkshire from the NEXT Map Digital Terrain Model (DTM) which provides a five metre horizontal resolution and one metre vertical accuracy. A DTM is a type of DEM which represents the earth's surface without the presence of vegetation, buildings and other human engineering features. The NEXT Map DTM has been derived from airborne Interferometric Synthetic Aperture Radar (IFSAR) (http://www.neodc.rl.ac.uk), which allowed large areas of Britain to be rapidly covered at a high resolution.

The visualisation of geomorphological features has been performed using simulated solar shading at 315°and 45° to enhance features. A single illumination angle can result in some features parallel to the illumination azimuth being suppressed (Clark and Meehan 2001). The map was then constructed on the recognition of features by size, shape, colour and tone according to a landform classification scheme using on screen digitisation. For example, a moraine ridge may appear as a linear raised area of topography and has been drawn as a brown line marking its extent. In addition other linear, arcuate and sinuous features are highlighted in brown and recognised by a raised area. Moraines are considered to be larger areas of raised land that are softer than the hilly surrounding, marked in pink. Finally glacial lakes are considered to be large areas of flat un-featured land, marked in blue. However, this relies on the assumption that all landforms have distinct characteristics (Hubbard and Glasser 2005). In addition, in complex bedform areas there can be considerable scope for misidentification as a result features overlaying one another or varying orientations. To supplement the mapping from the DEM the published BRITICE data (Clark *et al.* 2004a) has also been overlaid onto the map. BRITICE is the collection of previously published glacial features which have been complied onto a GIS data base forming a map of the glacial features of Great Britain. These features include glacial lakes, eskers, drumlins, moraines and meltwater channels. The overlaying of these features has allowed a checking of landforms, along with the addition of features identified as a result of their sedimentology rather than geomorphology, for example some glacial lakes, or those difficult to identify in complex areas. All areas considered to be glacial lakes as part of the mapping process are double checked with BRITICE as the area may be flat lying for another reason. If a more accurate geomorphologic map is required then it would be necessary to visit the field site to verify the DTM.

3.7 Chronology

In order to construct a chronology that the glacial sediments of Upgang can be correlated to, a series of published dates have been used. Radiocarbon, thermoluminescence and cosmogenic dates were taken from published accounts of the last deglaciation from around Great Britain and Ireland. These included dates of readvances and of deglaciation of an area. For consistency those dates that have been published in radiocarbon years have been calibrated using Calib 5.0.1 (two sigma ranges). If however, calibrated years are published then these have been the dates reported within this thesis. Throughout this thesis, where possible, dates cited in the text are all calibrated, although some radiocarbon dates were too old to calibrate. In addition, tables include both the published radiocarbon and calibrated dates.

In addition to using published dates to create a chronology the process of wiggle matching vegetation records to the Greenland ice cores has been used. Studies have previously shown that there is a good match between the Greenland ice records and British climate (Lowe *et al.* 1995). This has allowed vegetation and climate records with little or no dating to be matched to a dated record.

4. Results and Analysis

4.1 Section Logging

The overview log of the cliff shows four distinct lithofacies associations (LFAs) within the cliffs; D1, D2, SG and D3. The lowest LFA (D1) is a grey clast supported diamict, separated from the overlying brown matrix supported diamict (D2) by a very distinct boundary. D1 is composed of brecciated bedrock that is highly compact. D2 is a relatively structureless diamict that has a moderate number of clasts. This LFA is possibly spilt into two; towards the north of the section a colour and clast content distinction separates the darker, clast rich D2A from the lighter D2B with a moderate clast content, although this could be a result of moisture content. The size of the larger clasts also fines upwards. Towards the south D2 appears to have a lighter colour with a stratified facies, D2C, below the homogenous diamict. Unlike the potential two facies to the north the stratified facies is clast poor in comparison to the overlying diamict and has a lighter colour. A sand and gravel layer (SG), which is subdivided into four facies, can then be seen across the entire cliff capped by a red clast poor diamict (D3). D3 is clast poor in comparison to D2, with smaller clast sizes and appears to have a finer grain size composition than the lower diamict and has a red tint to it.

Each LFA can be found in distinct layers throughout the section maintaining the sequence described above. No large scale faults or folds are seen. Several sections of the cliffs have also been logged at a smaller scale to provide greater detail; the location of each section is shown in Figure 6. This lithofacies found within these logs have been depicted on each Figure.

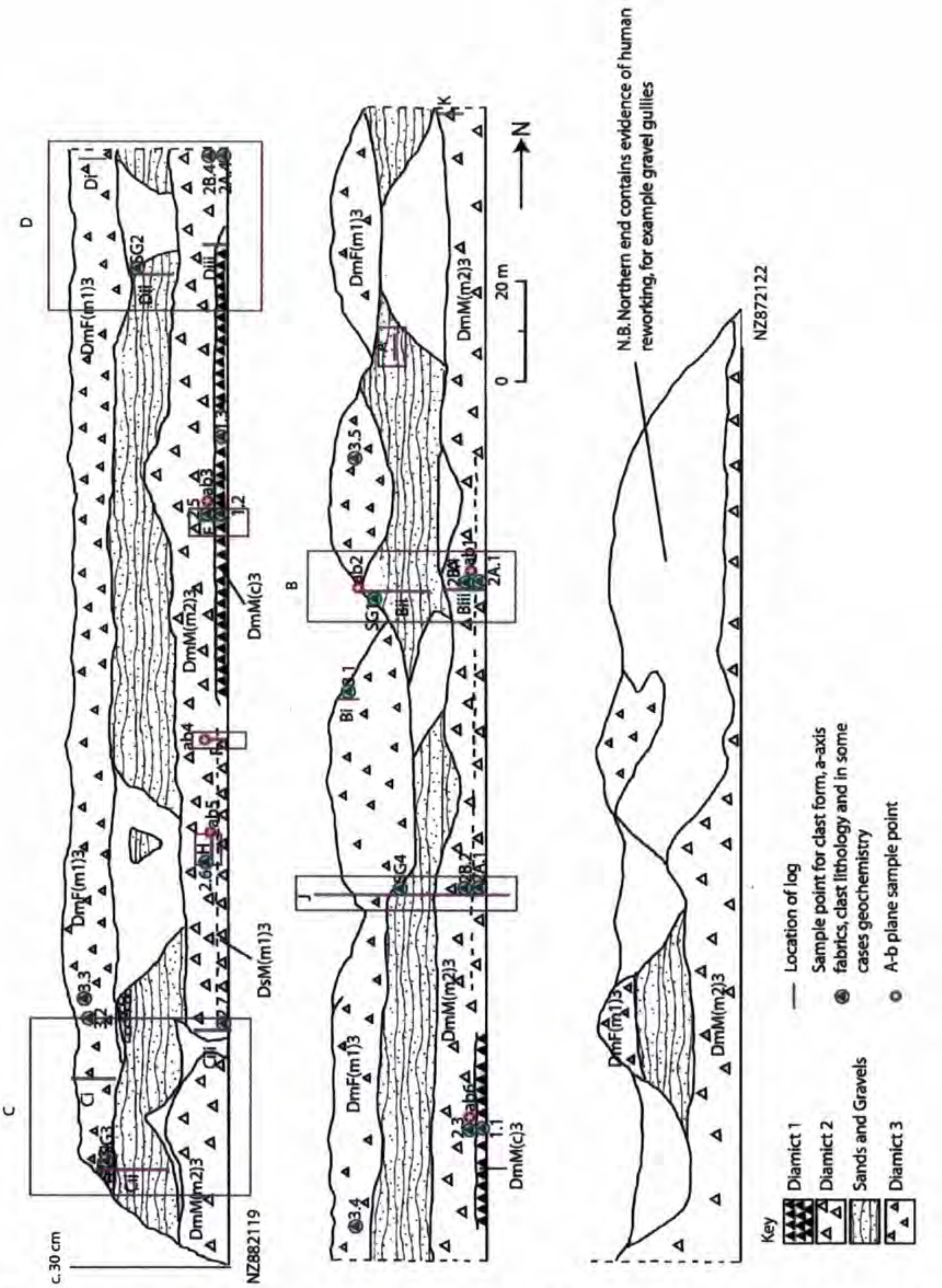


Figure 6: Overview of the cliff section showing each LFA and the sample sites. Logs are shown with lines representing the location of each section and a box indicating the general location of the site for clarity.

4.1.1 Section A

Section A (Figure 7) displays sub horizontally stratified sand and gravel. The section shows a series of laminated sands and granules at the base. To the left of the section there are planar laminated sands overlain by horizontally bedded granules and planar laminated fines. Towards the middle of the section these laminations become concave in nature angled up towards the fines unit. More concave, and some convex, laminations occur within this unit towards the right of the section in both sands and granules. The middle third of the diagram shows a unit of massive fines with a number of sand inclusions. Several of these lenses are oval or circular in shape (Figure 7). The sand within them tends to be laminated although the direction of these laminations varies between pods. The remaining lenses are elongate with concave upper and lower boundaries the sand laminations within these lenses display concave bedding. Above this unit lies an upward coarsening sequence of laminated sand and horizontally bedded gravels.

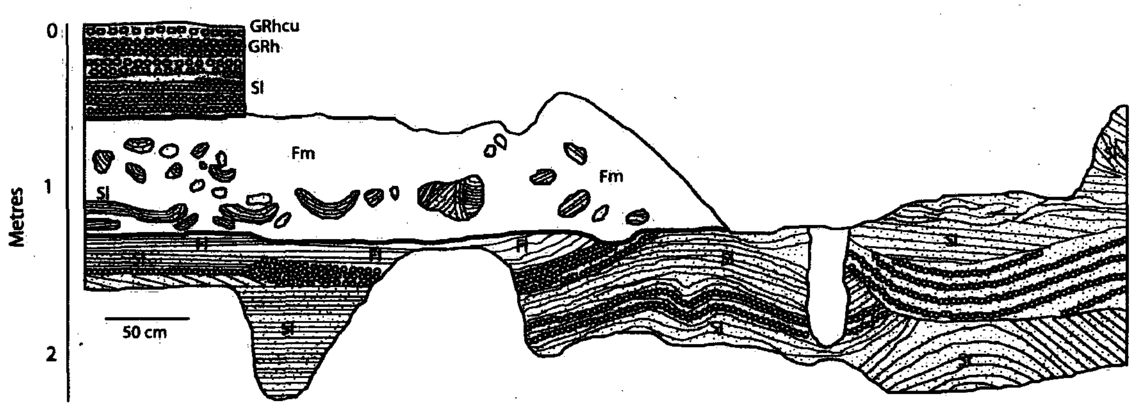


Figure 7: Log of Section A. This section is approximately 17 metres from the top of the cliff.

4.1.2 Section B

Section B logs the vertical profile of the cliff from D2 to D3 in three parts. The lowest Section Biii (Figure 8a) consists of homogenous brown diamict, D2. Potentially this LFA can be spilt into two separate facies, D2A and D2B. The lower 80 cm to 1 metre of the section, D2A, is darker in colour (7.5YR 4/3) with a greater number of clasts of a larger size than the over lying diamict, D2B (7.5YR 5/3). The two facies are separated by a gradual boundary, with the differences in colour as the major difference. D2 is then overlain by a complex sequence of sands and gravels (Figures 9a and 9b). The lower part of this sequence, SG1, is dominated by finer sediments (clay, fines and fine grained sand) which are mainly planar laminated with some Type B climbing ripples towards the top of the facies. This then coarsens upwards, to bedded sands and fines, SG2, with Type B ripples within the sand. These ripples have paleocurrent directions to the north west and west, dipping at approximately 20°. The upper section consists of planar bedded coarse sand, granules and gravels, SG3. Within the gravel facies pods of sand and trough bedded sands can be found. These sand pods are broadly oval in shape with no pinching at either end and sharply defined boundaries with the surrounding sediments. The majority of them consist of laminated sands. The laminated trough bedded sands have a concave lower edge and a more horizontal upper boundary. At the very top of the LFA, SG4 represents a facies of sand with Type A ripples within it. This facies is very thin; approximately 20 to 30 cm. A relatively sharp boundary with the overlying diamict exists, although there appears to be some rafts of diamict within the sands. These rafts are reddish brown the same as the overlying D3 (7.5YR 4/6). They have distinct, angular edges with no mixing with the surrounding sand visible. The sand sequence to the right of the rafts appears to have been disturbed. The laminations appear as though they have been compressed upwards creating a slight wave in the laminations. Paleocurrents from this section can be seen on the diagram and tend to be from between NNW and NNE with dip values of approximately 22 °. Figure 8b shows the final section of the cliff Bi. This is D3, a reddish (7.5YR 4/6), clay rich, clast poor diamict with various inclusions. Towards the bottom of this section a large section of sands has been included within the diamict consisting of laminated sands, fines and granule bands, which is only partly exposed. This feature has a strongly defined horizontal upper boundary with the overlying diamict; the other edges of the feature are obscured. The laminated sand and fines within this block show some evidence of a fault at the base of the feature. The next feature is a sand pod, consisting of lighter golden coloured sand (10YR 4/3), with fine coal particles. The pod itself has been pinched at either end, forming an 'eye' shaped feature. This sand contains some internal laminations, although is dominantly massive with small granule sized clast towards the outer edges of the feature. Above this feature is an inclusion of reddish (5YR 5/6) clay and fines which has been pinched at one end, as well as horizontal reddish silt/sand bands (5YR 5/6).

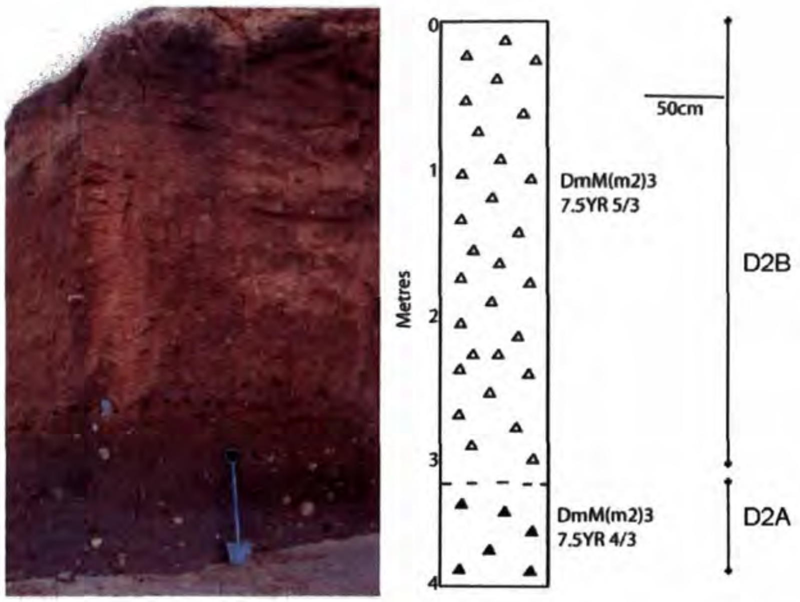


Figure 8a: Logs and photographs of the diamicts from D2 within Section B. Vertical scale from top of section approximately 19 meters from top of cliff. Lithofacies associations shown to right of diagram.

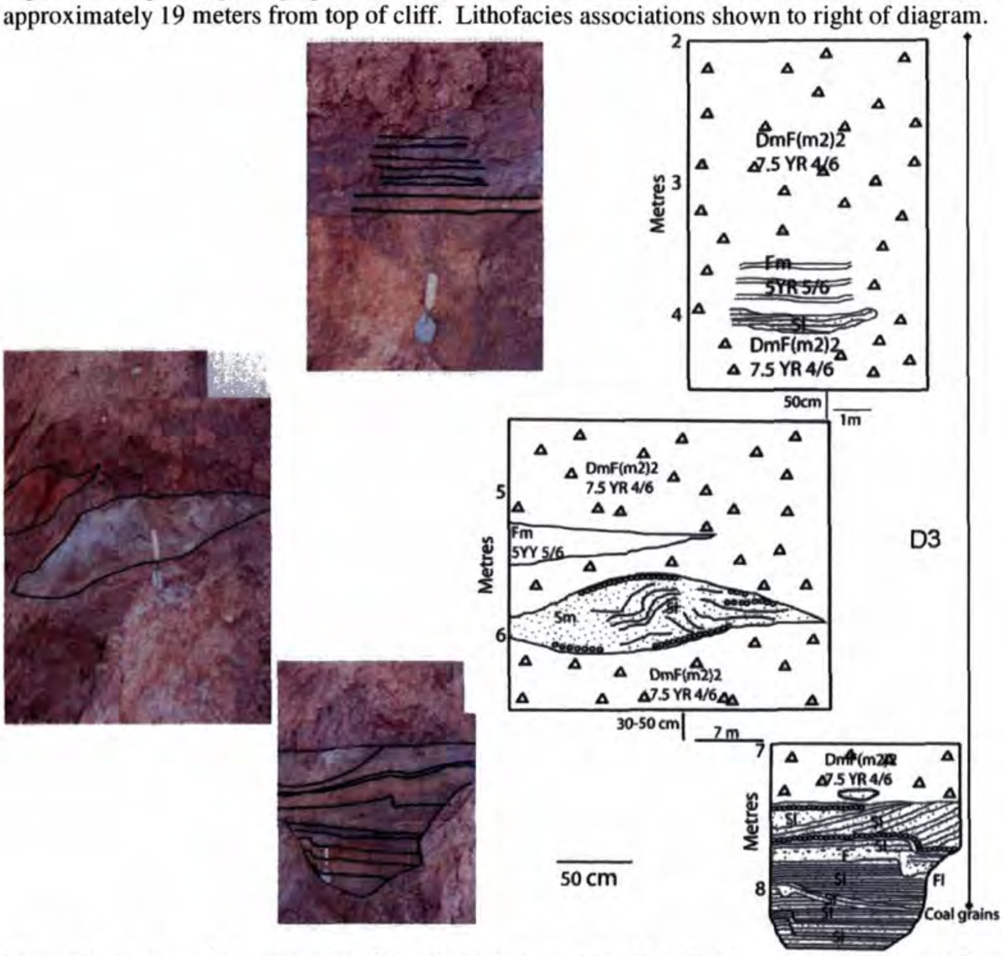


Figure 8b: Logs and photographs of the diamicts from D3 within Section B. Vertical scale from top of cliff. Lithofacies associations shown to right of diagram.

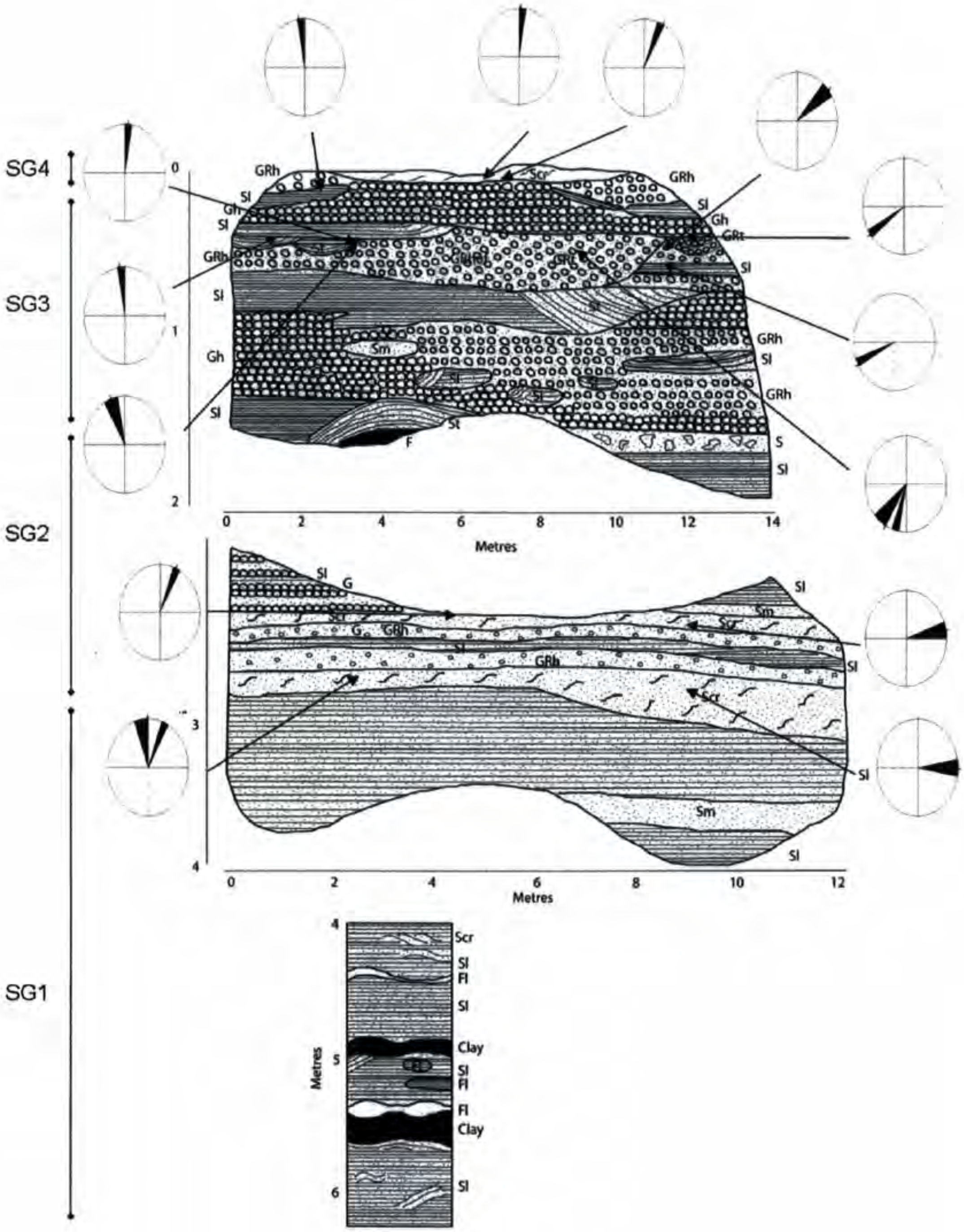


Figure 9a: Logs from the sand and gravels LFA (SG) in Section B. The individual SG facies have been named to the left of the diagram. Vertical scale from top of SG section not top of cliff. SG section approximately 10 to 11m from top of cliff.

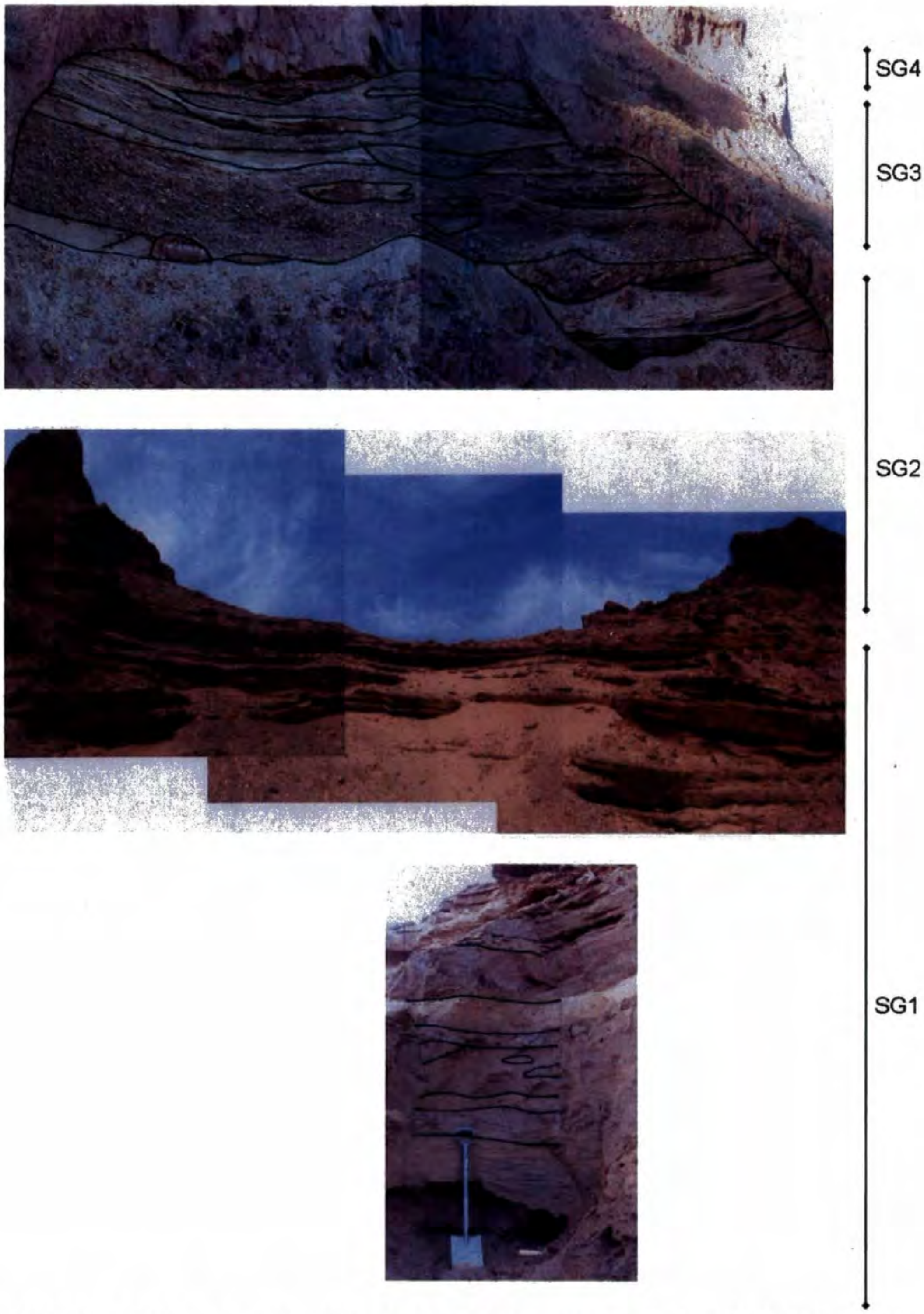


Figure 9b: Photographs from the sand and gravels LFA (SG) in Section B. The individual SG facies have been named to the right of the diagram. Spades for scale. SG section approximately 10 to 11m from top of cliff.

4.1.3 Section C

Section C is another vertical profile of the section (Figure 10). The lower LFA, D2, can be split into two facies. At the base of the section is D2C, a light brown (7.5YR 3/2) diffusely stratified diamict. D2C is a lighter colour with less clasts than the overlying homogenous diamict of D2. The sands and gravels overlie D2, although the boundary between diamict and the sands is not visible. This sand and gravel sequence follows a similar pattern to the one logged at Section B. At the base of the section is homogenous brown clay overlain by laminated sands and fines (SG1) with some Type B climbing ripples. This is followed by SG2 which is a section of largely massive sands with some Type B climbing ripples. This facies incorporates a clay inclusion, narrow at one end widening to the right of the diagram with a semi-circular end to it. The edges of this feature are clearly defined from the surrounding sand with no mixing of the two sediments. Again the sequence coarsens upwards to gravels, SG3, although it was not possible to log this section in detail. At the top of sequence is D3 which is capped by topsoil. D3 is the reddish (7.5YR 4/6), clast poor clay rich diamict. This section also has sand and fines inclusions within it. The lower features are laminated sand rafts at approximately 35° to the horizontal. The sand is golden in colour (10YR 4/3), as was the sand pod in Section B, and includes small coal particles. Two of these rafts are oval in shape while the other two are obscured in places. The visible ends of these rafts are not smooth and appear flame like, interlinking with each other, although the two rafts remain separate. Towards the top of the section is a laminated clay and sand inclusion. The clay appears as the same colour to the diamict with reddish sand. The feature is narrow to the right of the diagram widening towards the left before forking into two thinner bands. The sediment within the inclusion is finely laminated, with laminations following the shape of the feature.

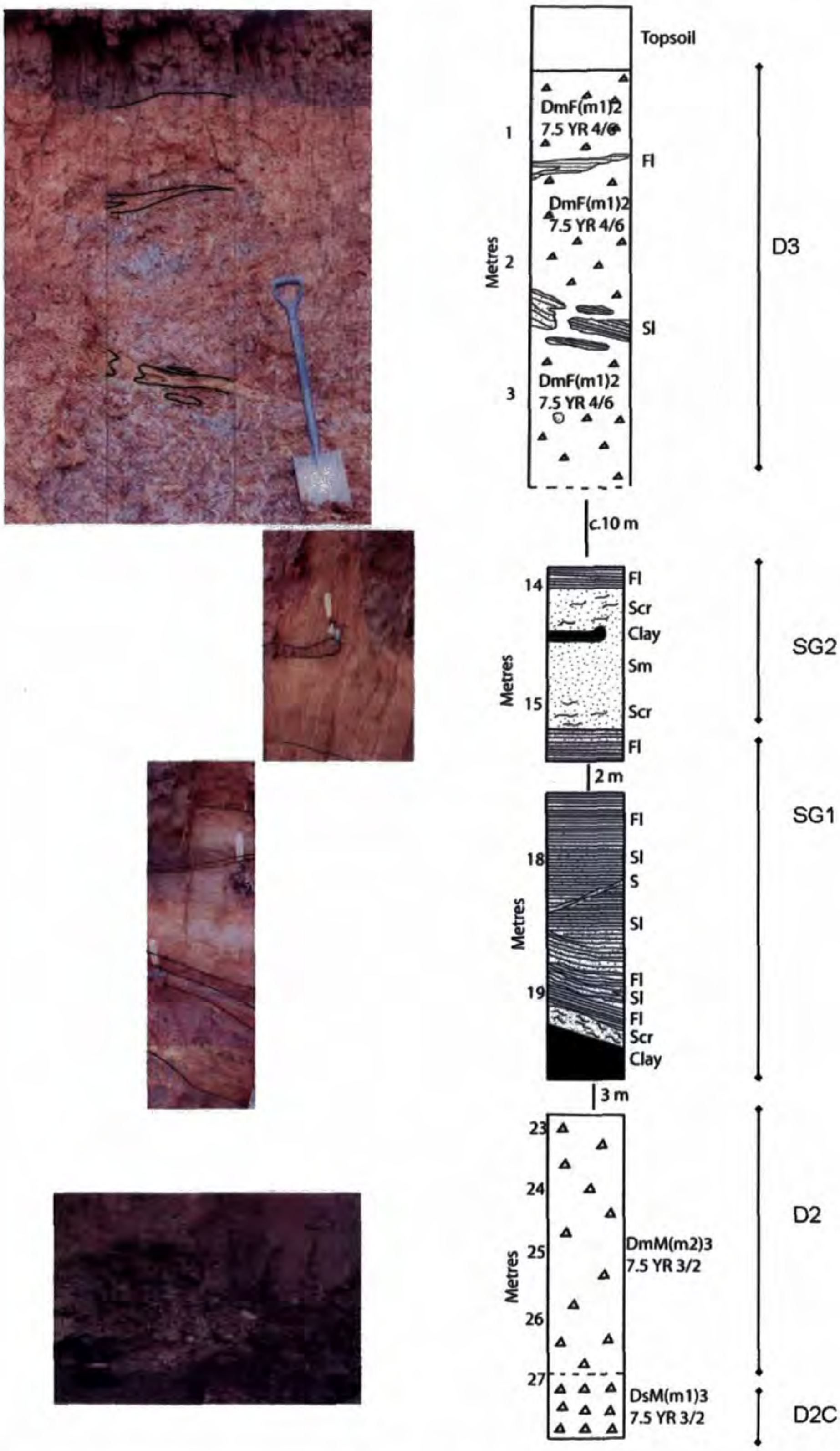


Figure 10: Logs and photographs from Section C. Spades and trowels for scale. Vertical scale from top of cliff. Each lithofacies has been depicted to the right of the diagram.

4.1.4 Section D

Section D shows a vertical log of the cliff including all four LFAs (Figure 11). At the base of the section is D1 which to the right of the section is overlain by a red diamict (7.5YR 4/3). D1 is grey in colour (10YR 5/1) consisting of very compact brecciated rock. The overlying red diamict is very clast poor and only outcrops for a short distance in a wedge shape between D1 and D2, wider at the base narrowing upwards as it arcs over D1. The boundary between D1, the red diamict and D2 above are all very distinct with no mixing of sediment. However, two oval shaped inclusions of this red diamict can be found, one in D1 and the other directly above in D2. D2 in this section is homogenous and brown (7.5YR 3/2), although no distinction between D2A and D2B is seen. Above D2 is the sand and gravel LFA that can be seen throughout the cliff. Again the base of the section is characterised by planar laminated fines, SG1, grading up into alternatively bedded sand and gravel facies, SG3. Within this section SG3 is represented by a continuous switch between horizontally bedded gravels and planar laminated sands. There are no inclusions found within this sand and gravel LFA. The section is topped by D3. Within this section D3 is reddish in colour (7.5YR 4/6), clast poor and appears to have a high clay content. However, towards the top of the section there appears to be some mixing of the diamict with clay and silts. These clays and silts are red (5YR 3/1) and blackish (5YR 2/6). Mixing is incomplete with the various colours and sediment types seen. However, there are no clear boundaries between these sediments. This part of the section is very clast poor and the clasts found are small in size. The sequence is capped by topsoil.

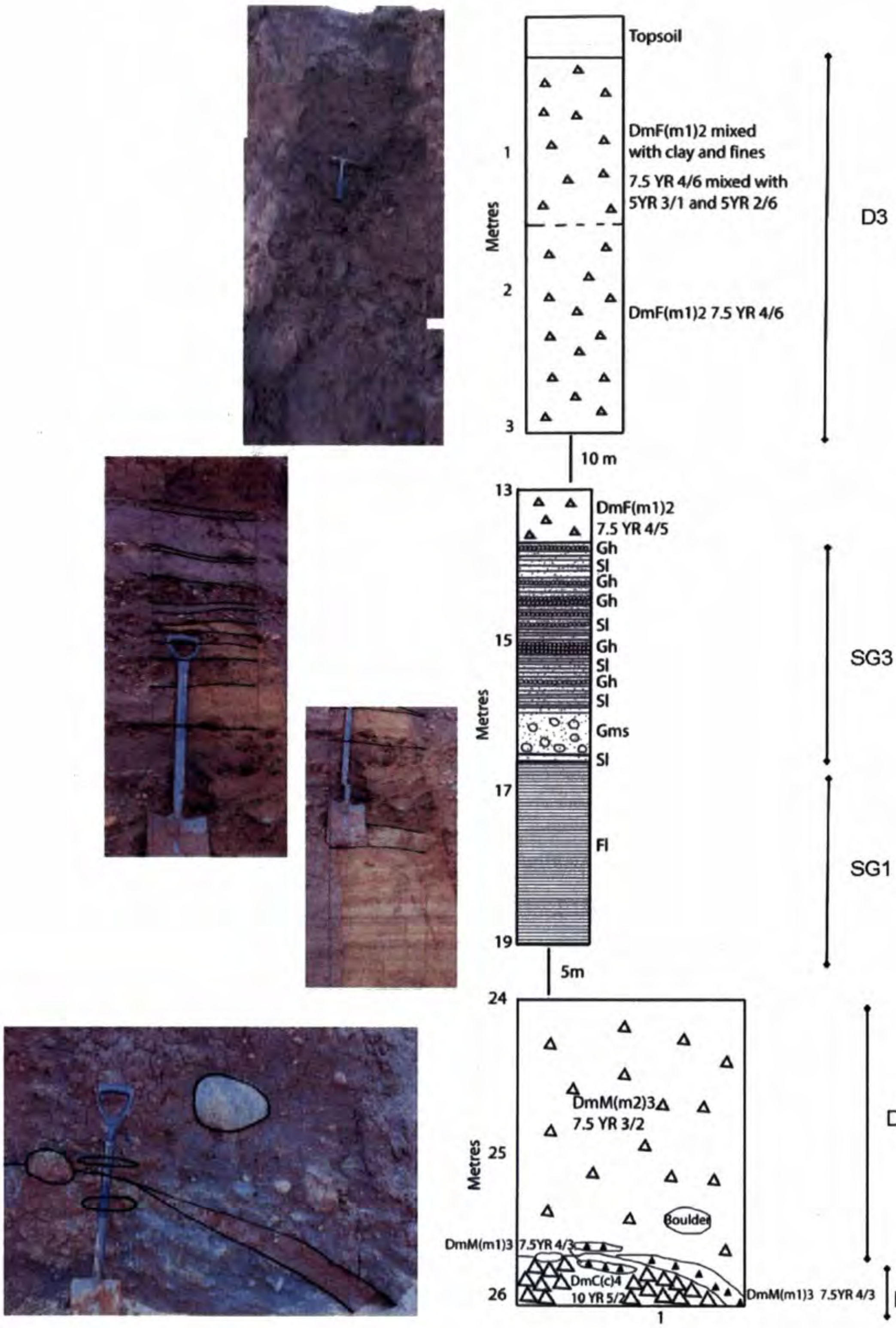


Figure 11: Logs and photographs from Section D. Spade and geological hammer for scale. Vertical scale from top of cliff. Each lithofacies has been depicted to the right of the diagram.

4.1.5 Section E

Section E focuses on the boundary between D1 and D2 (Figure 12). At this site D1 is a light grey colour (10YR 3/1) and is made up of very compact brecciated shale. There appears to be no structures or post depositional deformation present within the facies. The overlying D2 is a brown (7.5YR 3/2) matrix supported diamict. It has a relatively large number of granule to gravel sized clasts within it and appears relatively firm. The boundary between the two LFAs is very clear, sharp and distinct. There is no evidence of mixing or any erosional features. This sharp boundary can be seen between the two diamicts throughout the section.

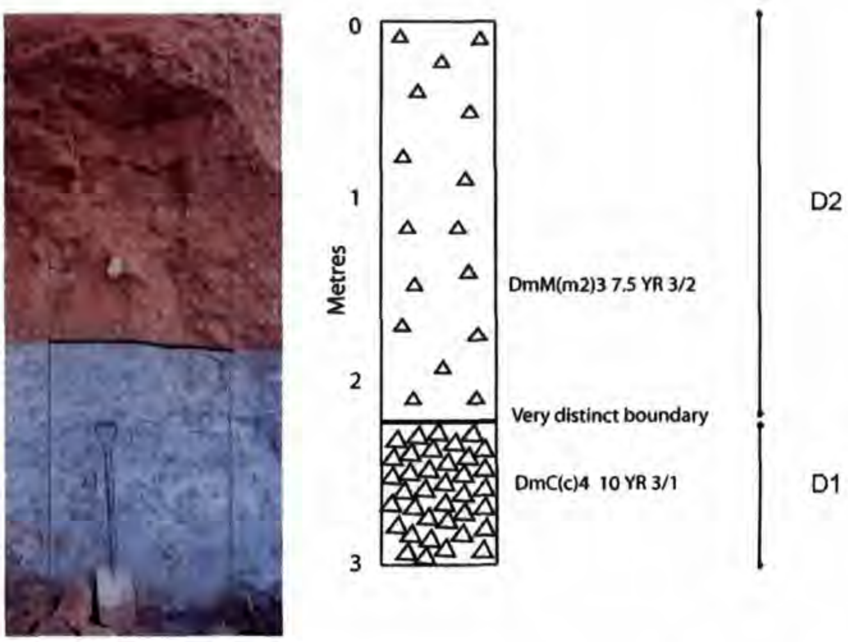


Figure 12: Photograph and log from Section E. Spade for scale. This section has been taken from the very base of the cliff. Each lithofacies has been depicted to the right of the diagram.

4.1.6 Section F

Figure 13 shows the inclusion of grey clast supported D1 into D2. This section of D1 is dark grey (10YR 3/1), clast supported brecciated shale. This has been classed as D1 on the basis of the colour, lithology and composition which are identical to this diamict. The underlying and overlaying D2 is a brown (7.5YR 3/2), relatively clast rich, homogenous matrix supported diamict. The attenuated wedge feature is pinched to the right of the diagram. However, the left side of the feature has been obscured. This general attenuation from left to right is interrupted in the middle section by a marginal indent. Distinct boundaries are found surrounding the feature separating the two diamicts. To the far right of the diagram the main inclusion of D1 is separated from a second inclusion of D1 by a thin oval shaped pod of massive sand. This secondary inclusion of D1 appears to be widening towards the right from its pinched end. Within the clast supported grey diamict are two elongated planar inclusions of massive grey fines. While below the main feature, at the pinched end is a thin horizontal band of reddish sand/silt.

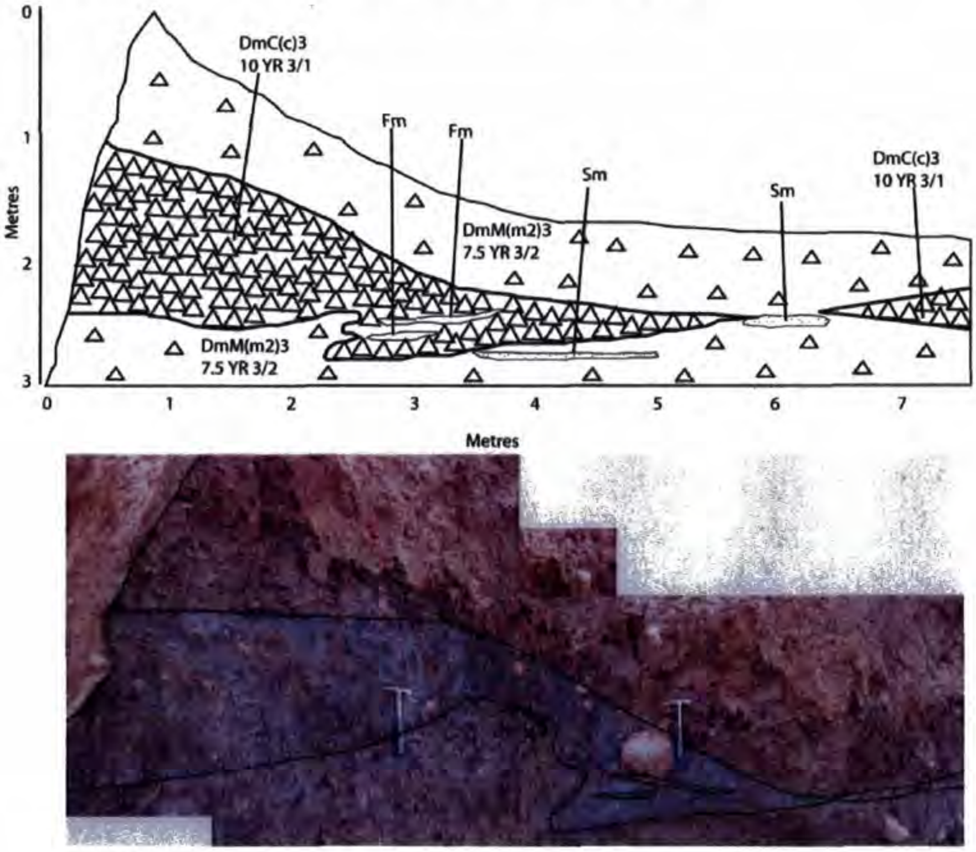


Figure 13: Log and photograph from Section F. Geological hammers for scale. The base of this feature is c. 50 cm from beach level.

4.1.7 Section H

Another inclusion within D2 is shown in Figure 14. This feature is an elongate, attenuated pod of orange fine sand or silts within the brown diamict. It has formed flat lying lenses with flame like structures. Within the middle and to the left are inclusions of fine golden brown sands. The sand inclusion to the left of the fines has also been attenuated at either end while the shape of the inclusion in the middle reflects the surrounding fines. Below this fines and sand feature are three smaller pods of dark grey (10YR 3/2) coarse sands.

Two of these pods are oval to circular in shape, while the pod furthest to the left on the diagram has a shape similar to two diamonds joined together.

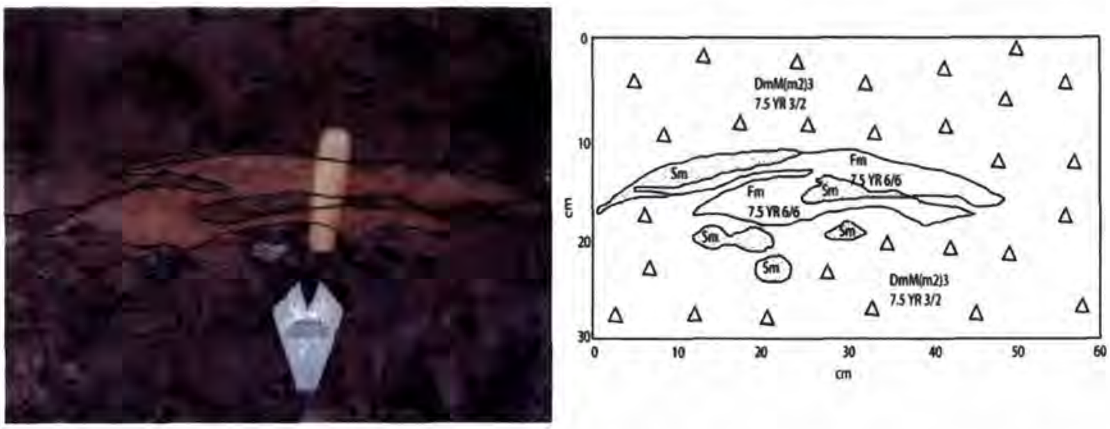


Figure 14: Photograph and log from Section H. Trowel for scale. This section is c. 2 m from beach level.

4.1.8 Section J

Section J is another vertical section (Figure 15) which focuses on the sand and gravel section. At the base of the section is D2, which can be separated into D2A and D2B, according to colour and clast content. The overlying sand and gravel section shows a similar pattern to those discussed in other locations along the section. The lower part of the study area consists of planar laminated sand and fines, SG1. This facies grades into facies SG2 which consists largely of planar laminated sands, with some Type B ripples, some foreset bedding and a gravel layer. These ripples have an orientation of SW to W with dips between 16 and 30°. Paleocurrent directions taken from the foreset bedding also have a SW to W orientation with a dip of 28°. Also within this facies is a pod of matrix supported gravel. This pod is elongate in shape, angled approximately 40° from vertical and lies through boundaries between the surrounding sand units. This facies is then overlain by coarsening up gravels. The gravels at the base of this facies are massive with a concave upper boundary to the unit. A thin fines band, SG4, separates the gravels from the overlying D3.

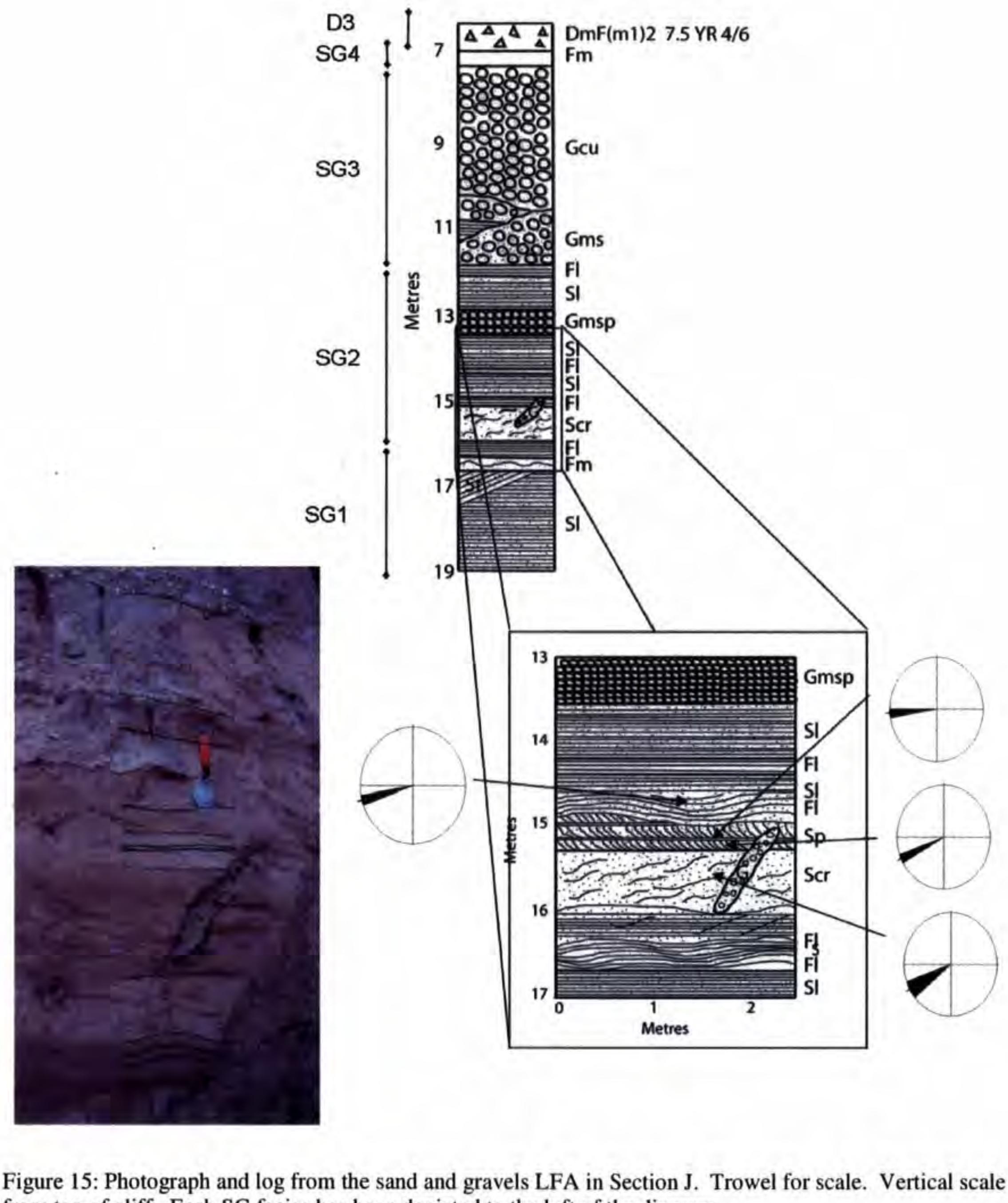


Figure 15: Photograph and log from the sand and gravels LFA in Section J. Trowel for scale. Vertical scale from top of cliff. Each SG facies has been depicted to the left of the diagram.

4.1.9 Section K

Section K shows an isoclinal fold nose of darker diamict within D2 (Figure 16). At this site D2 is a light brown (7.5YR 5/3) homogenous massive diamict. It is relatively clast rich with clast sizes appearing to fine upwards. The darker band that forms the fold nose is a similar colour to D2A (7.5YR 4/3). It is clast poor, clasts found are granule sized, opposed to the gravel sized clasts found in the surrounding D2. The fold itself is angled with the lower limb at a much steeper angle than the near horizontal upper limb. This darker band is potentially the same facies as D2A which has been included into D2B.

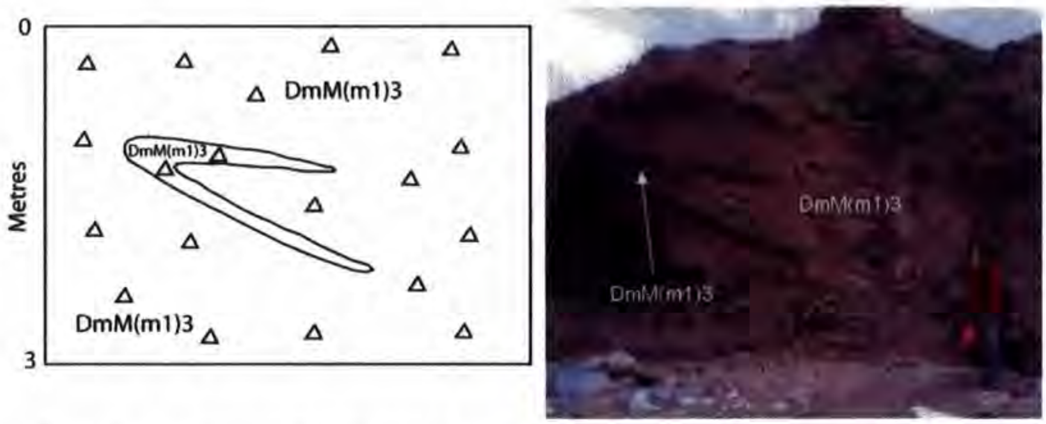


Figure 16: Log and photograph from Section K. Person for scale (standing at beach level).

4.2 Clast Form

4.2.1 Ternary Diagrams

Ternary diagrams have been drawn up to represent the shape of the clasts sampled (Figure 17). These diagrams plot the c:a ratio against the b:a ratio representing the clast shape with block like clasts plotting at the top apex, slabs towards the bottom left apex and elongate clasts at the right apex. The diagrams representing data from D1 show that the clasts are compact and block like, with the majority of points plotting above the 0.4 line. The points appear to be evenly spread along the horizontal representing platy, blade and elongate clasts without preference. The diagrams representing D2 are evenly scattered throughout the diagrams. However, there are no points in the extreme corners of the triangles. There is no pattern to distinguish between samples belonging to D2A, D2B and D2. D3 shows a similar pattern to the samples in D2, although the points plot higher up the triangles.

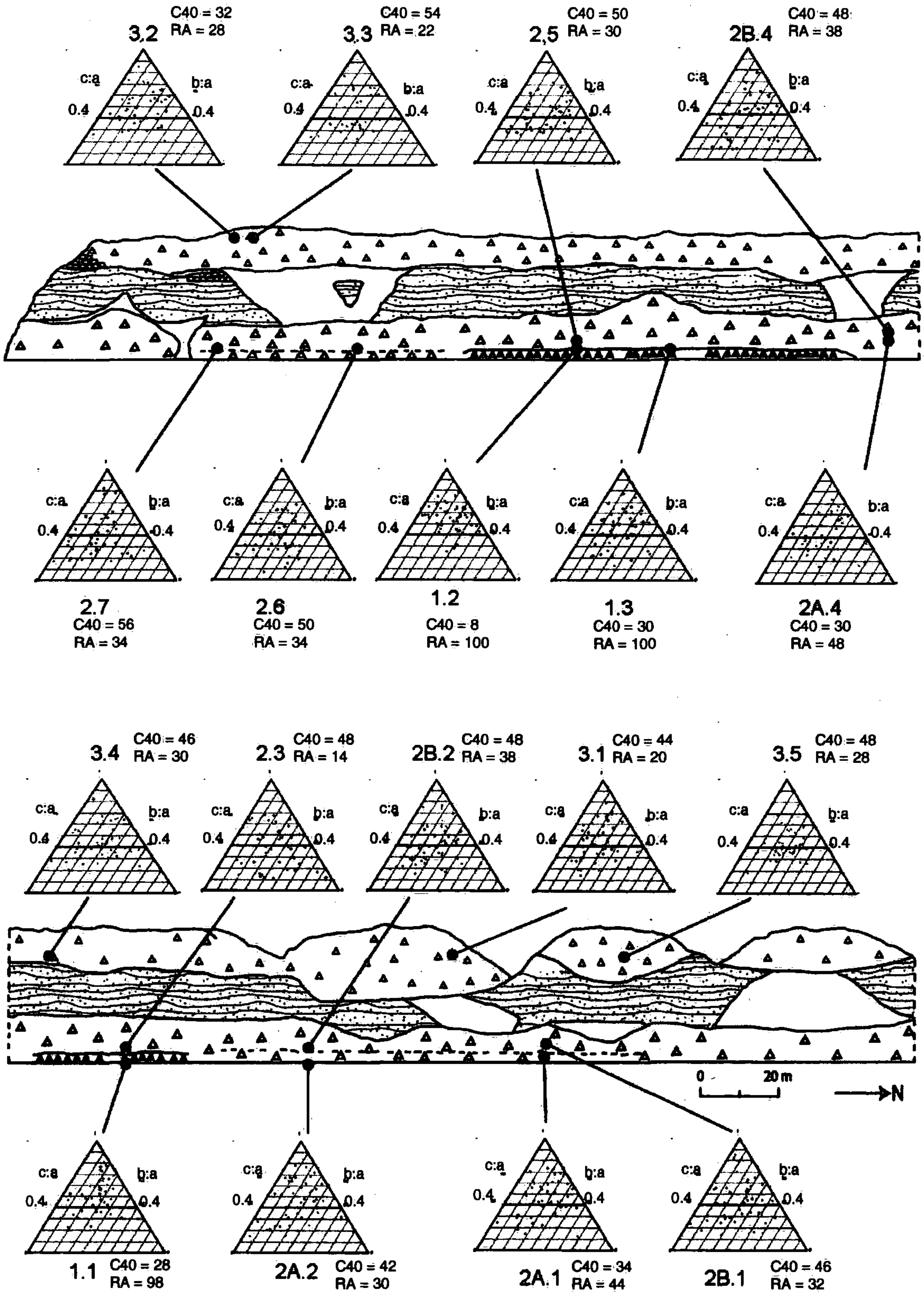
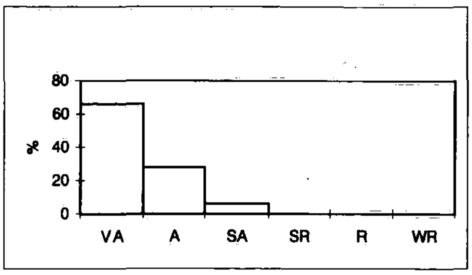


Figure 17: Clast form triangles from Upgang and the location of sample.

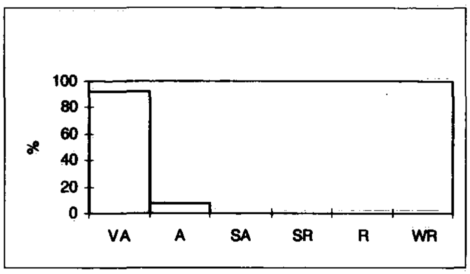
4.2.2 Roundness histograms.

Histograms created from the roundness data area shown in Figure 18. Samples taken from D1 have a high percentage of very angular clasts, over 65%, with the remaining clasts classified as angular. The percentage of very angular clasts compared to angular clasts varies according to lithology; those with a higher percentage of very angular clasts have a higher percentage of shale clasts. Data taken from the remaining diamicts appears very similar, with a high percentage of sub-angular clasts, between 35% and 65%. The samples from D2 show a greater variability, whereas the percentages of sub-angular clasts from D3 are consistently 50-55%. The percentage of angular clasts in each sample ranges between 20% and 45 %. Two samples from D2A show higher values in angular clasts with 2A.1 and 2A.4, 40% and 45% respectively. However, the remaining sample from D2A has a relatively low value of 25%, meaning that there may be some variation throughout the cliff laterally as well as vertically. Rounded clasts also have a significant percentage ranging between 10% and 25%. There appears to be no significant correlation between the location of samples and the percentage of rounded clast

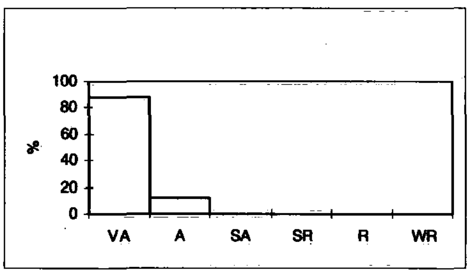
Site 1.1



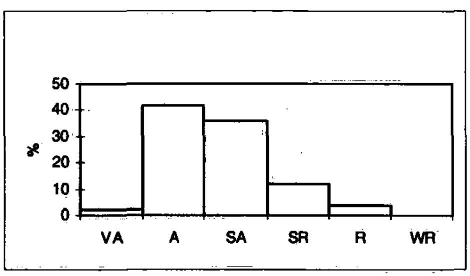
Site 1.2



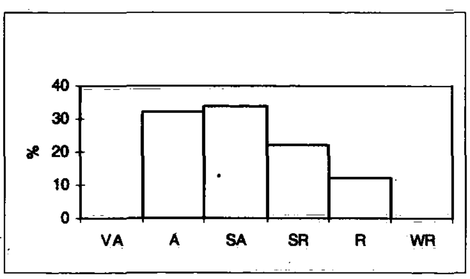
Site 1.3



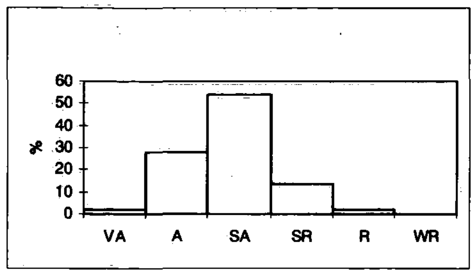
Site 2A.1



Site 2B.1



Site 2A.2



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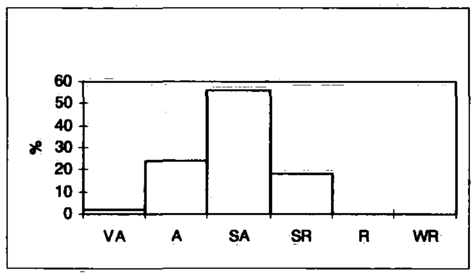
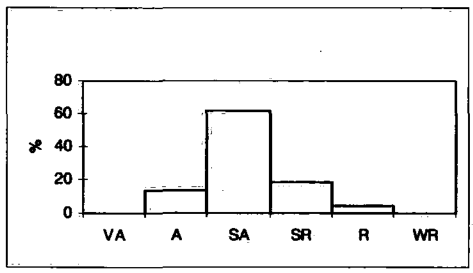
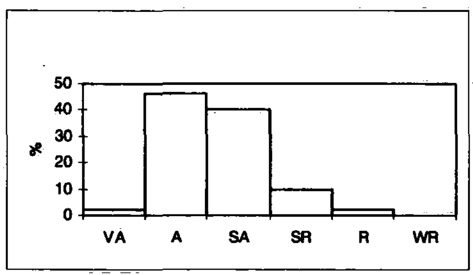


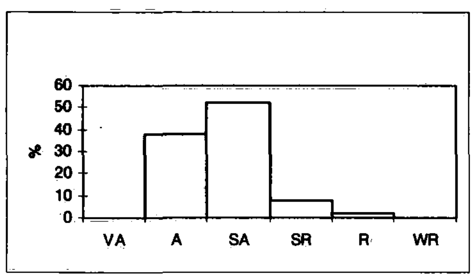
Site 2.3



Site 2A.4



Site 2B.4



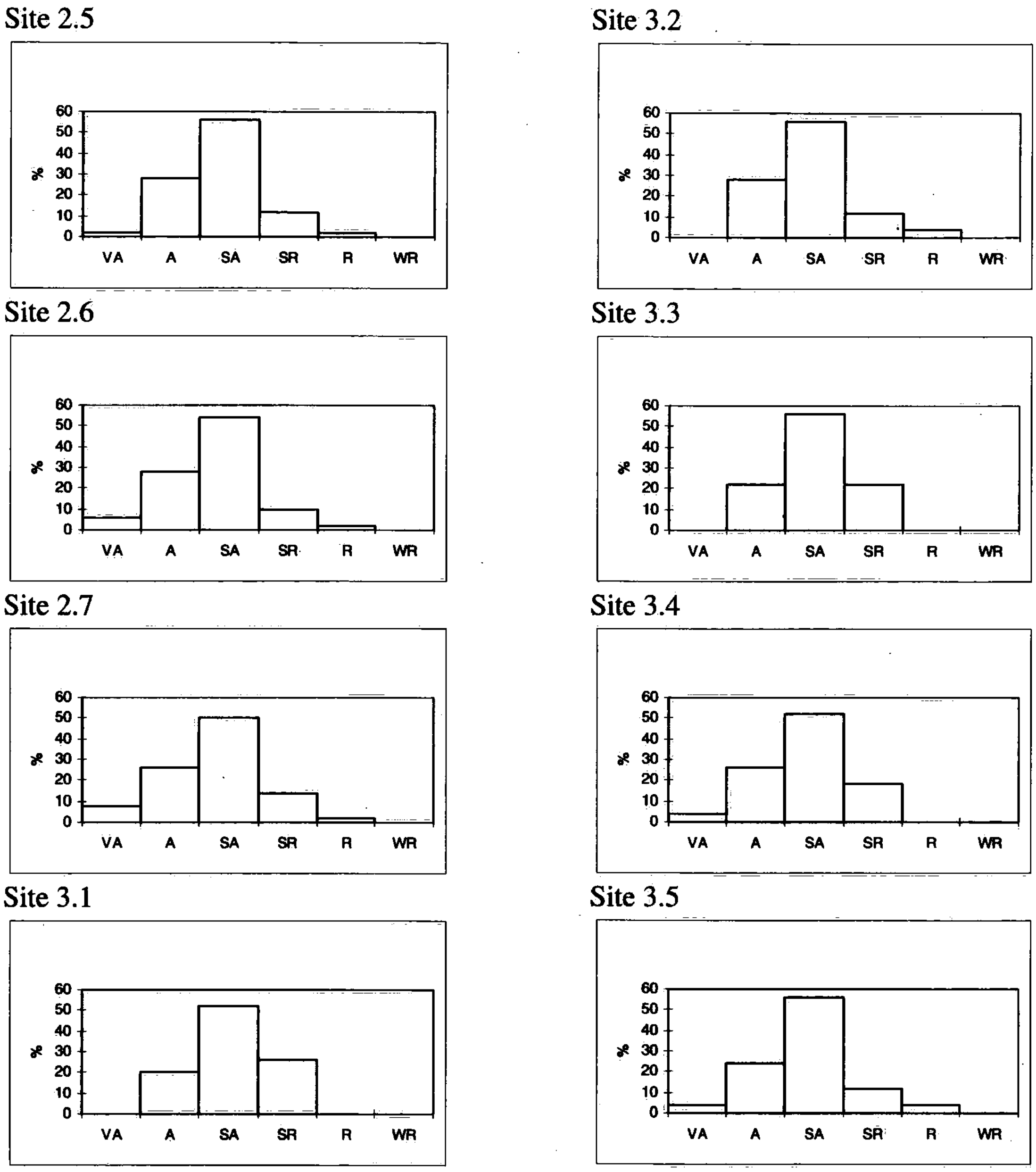


Figure 18: Roundness histograms from Upgang. Location of each sample can be seen in Figure 6.

4.2.3 Covariance Plot

Figure 19 shows the covariance plot created from the data colleted at Upgang, plotting C40 (shape) against RA (angularity). The points have been separated in colour depending on which diamict layer it belongs to. As a control it is expected that scree samples would plot in the top right corner with fluvial or purely subglacial material plotting in the bottom left corner.

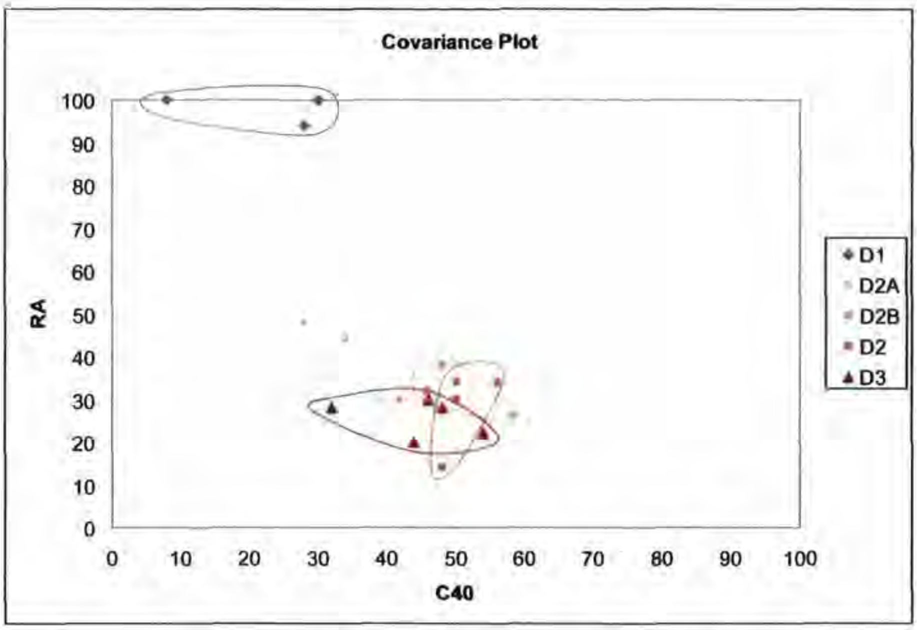


Figure 19: Covariance Plot indicating the C40 and RA of each sample site. Each facies has been colour coordinated and circled to distinguish where each LFA plots.

The three samples from Diamict 1 plot towards the top left hand corner, showing a very high RA and a low C40 value. The upper diamicts, D2 and D3, plot in similar areas with C40 values between 30 and 60 and RA values between 20 and 50. However, differences occur between the various tills. D2A has higher RA and low C40 values compared with D2B. The two facies are clearly separated from each other on the graph. The remaining points from D2 tend to plot closely to those from D2B, plotting with the exception of one point within the curve bounding D2B. D3 plots lower on the graph than the points associated with D2. They have lower RA values but similar C40 values to those above, with the exception of one point they tend to plot below those from D2B.

4.3 Clast Fabrics

4.3.1 Rose Diagrams

The rose diagrams plotted from the orientation of the a-axis of the clasts shows a general NE to SW trend in direction (Figure 20 and Figure 21). This trend varies on the mean trend of 45° by c. 20-30 °, in both D2 and D3. However, not all sites conform to this trend. 2A.1 shows no real pattern while 2.5, 2.7, 3.2, 3.3 and 3.5 show a different trend from NW to SE. In addition, samples from D2B seem to show a stronger orientation that those from D2A.

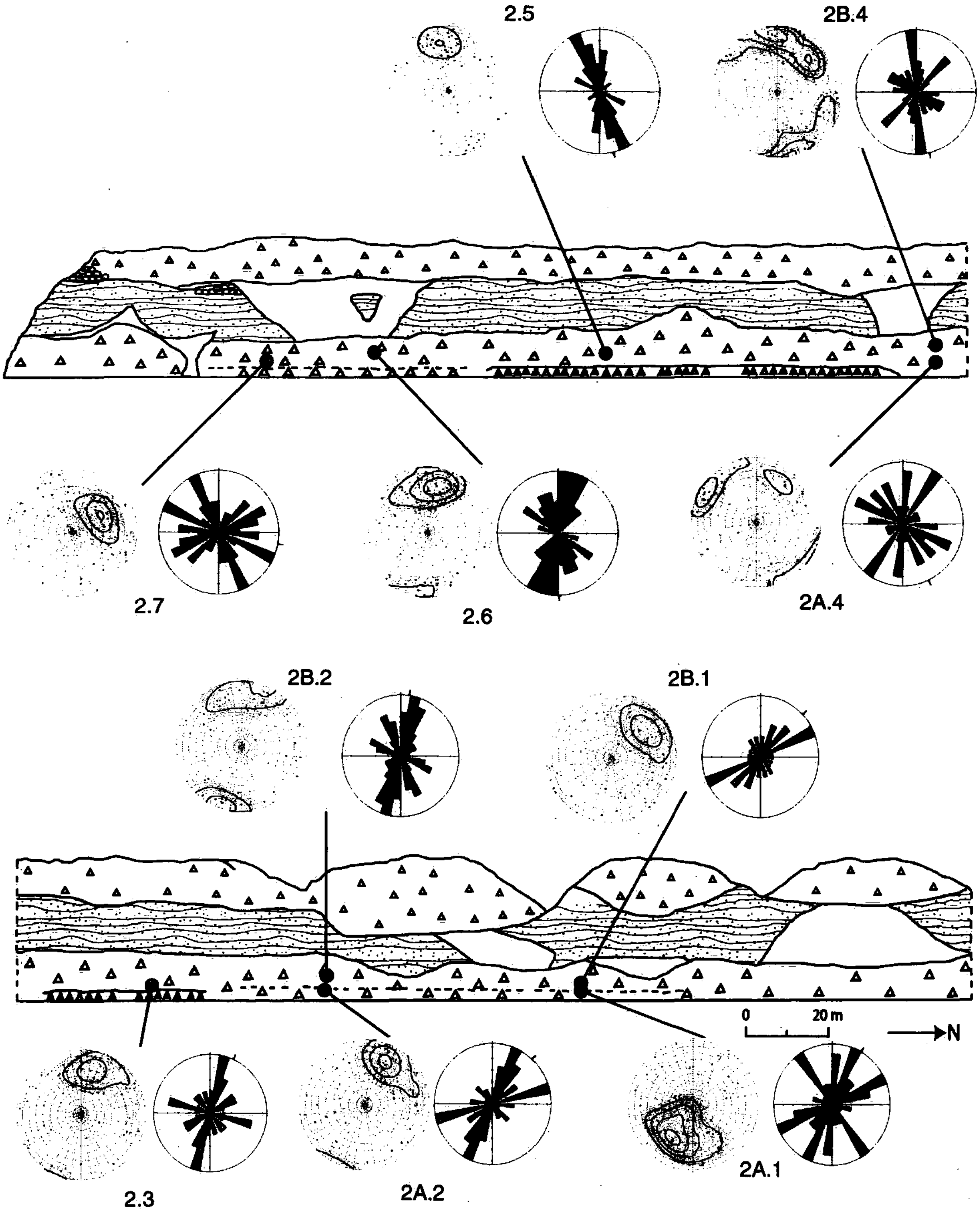


Figure 20: Rose diagrams and equal area stereonets from D2 (Gaussian Weighted contours).

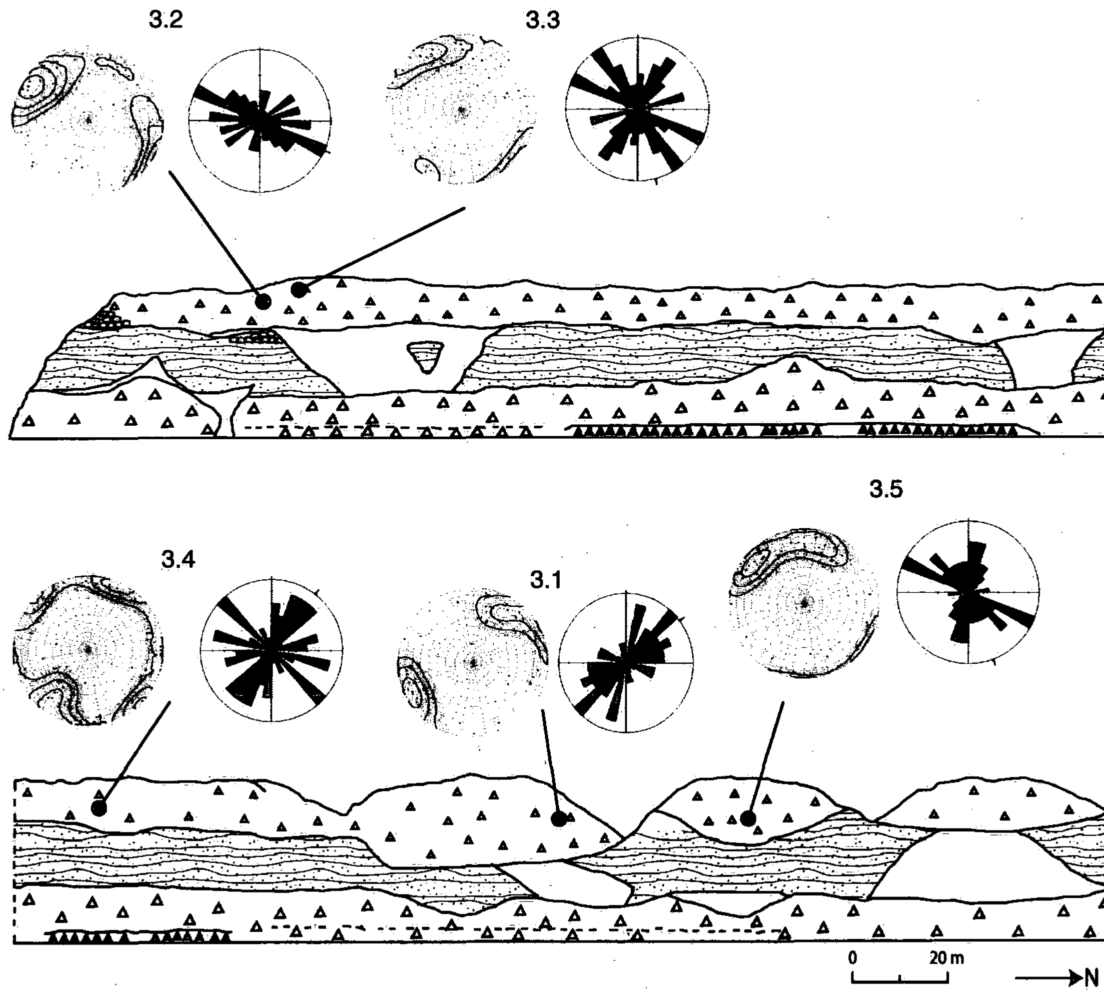


Figure 21: Rose diagrams and equal area stereonets from D3 (Gaussian Weighted contours).

4.3.2 A-axis equal area stereonets

Figures 20 and 21 also show the equal area stereonets plotted from the dip and azimuth data. These have been contoured using the Gaussian weighting method. These highlight the north east to south west orientation of clasts along with low dip values. Step function contours have also been calculated (Figure 22) so they can be classified on a five point scheme of modality (Hicock *et al.* 1996) as follows: Unimodal clusters (2A.2 and 2.5) meaning a tightly grouped cluster of a-axis points; Spread unimodal (2B.1 and 3.5) a single concentration of long axis points spread over a wider area; Bimodal clusters (2B.2 and 3.1) two grouped clusters at 90° to each other; Spread bimodal (2A.4, 2.6 and 3.3) loosely grouped bimodal concentrations and finally polymodal (2A.1, 2B.4, 2.3, 2.7, 3.2 and 3.4) consisting of multiple concentrations or a continuous spread around the perimeter. However, there does not appear to be a pattern between modality of each sample site and the lithofacies association that they originate from.

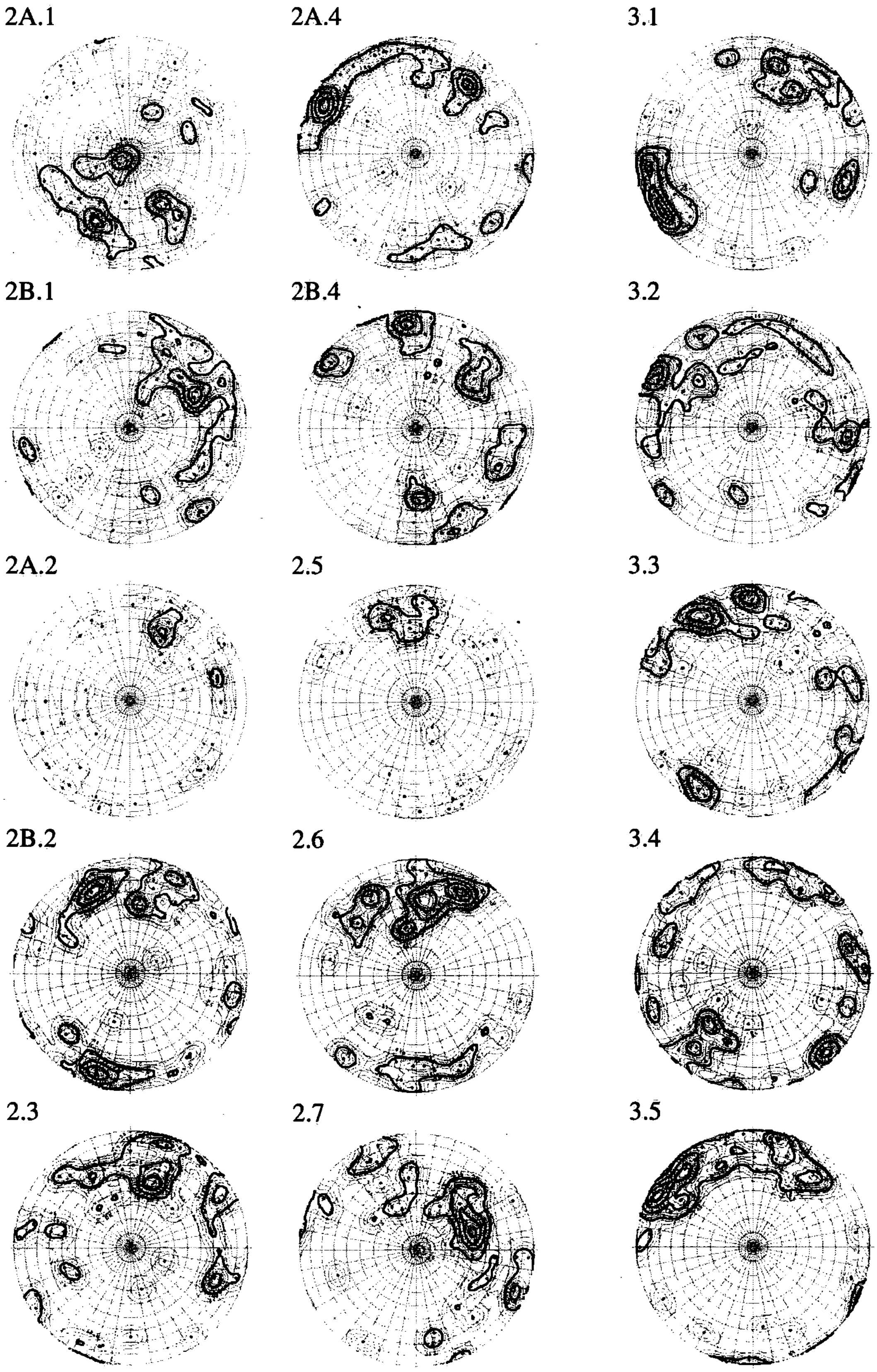


Figure 22: Step contoured equal area stereonets.

4.3.3 A-B plane rose diagrams and equal area stereonets

The rose diagrams constructed from the a-b plane fabrics (Figure 23) show either a dominant azimuth from NW to SE or NE to SW. When comparing these with those drawn using the a-axis data it is possible that the majority of a-axes orientated NE to SW are lying transverse to flow, as several a-axis diagrams also show the NW-SE trend. Alternatively the NE to SW trend agrees with the a-axis data. This pattern is again seen on the Gaussian weighted stereonet plots. However, the two fabrics taken from just above D1 do not show this same pattern. In site ab3 the direction of orientation is N to S, although there is considerable spread of points around the diagram while site ab6 shows a direction of E to W. The step function contour plots show that all the samples are polymodal.

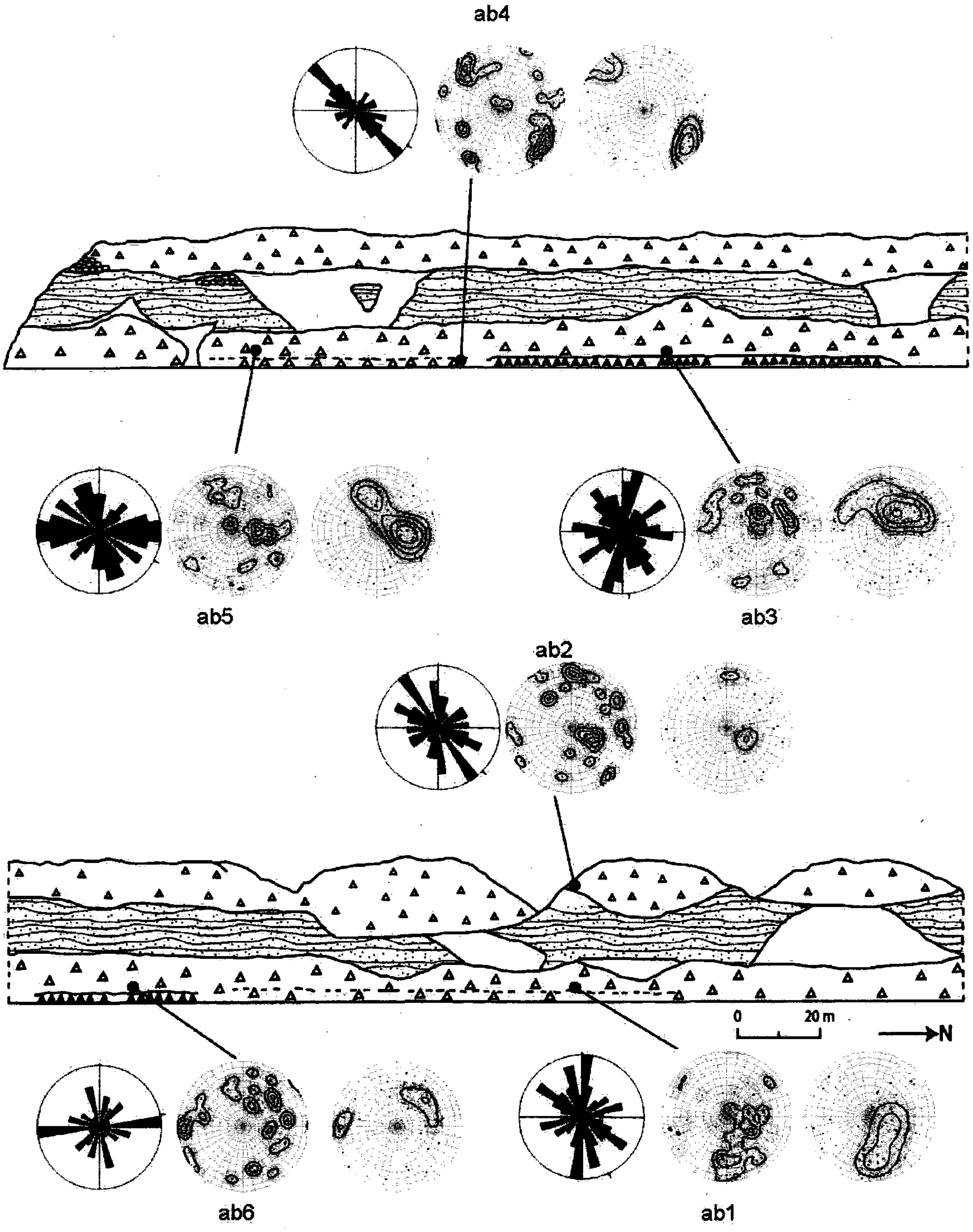


Figure 23: A-B plane data from Upgang. Rose diagrams, step contoured stereonets and Gaussian weighted stereonets respectively.

4.3.4 Eigenvalues plots

The eigenvalues calculated from the clast fabric data have been plotted on both a ternary diagram (Benn 1994) and a S3/S1 plot (Dowdeswell and Sharp 1986) (Figure 24). The points within the ternary diagram largely plot towards the lower centre with a few outliers.

They plot closer to the cluster vertex of the triangle than the girdle or cluster. On the S3/S1 plot these points on the left of the diagram in the centre. Both diagrams show considerable variation between samples and fairly poorly clustered fabrics.

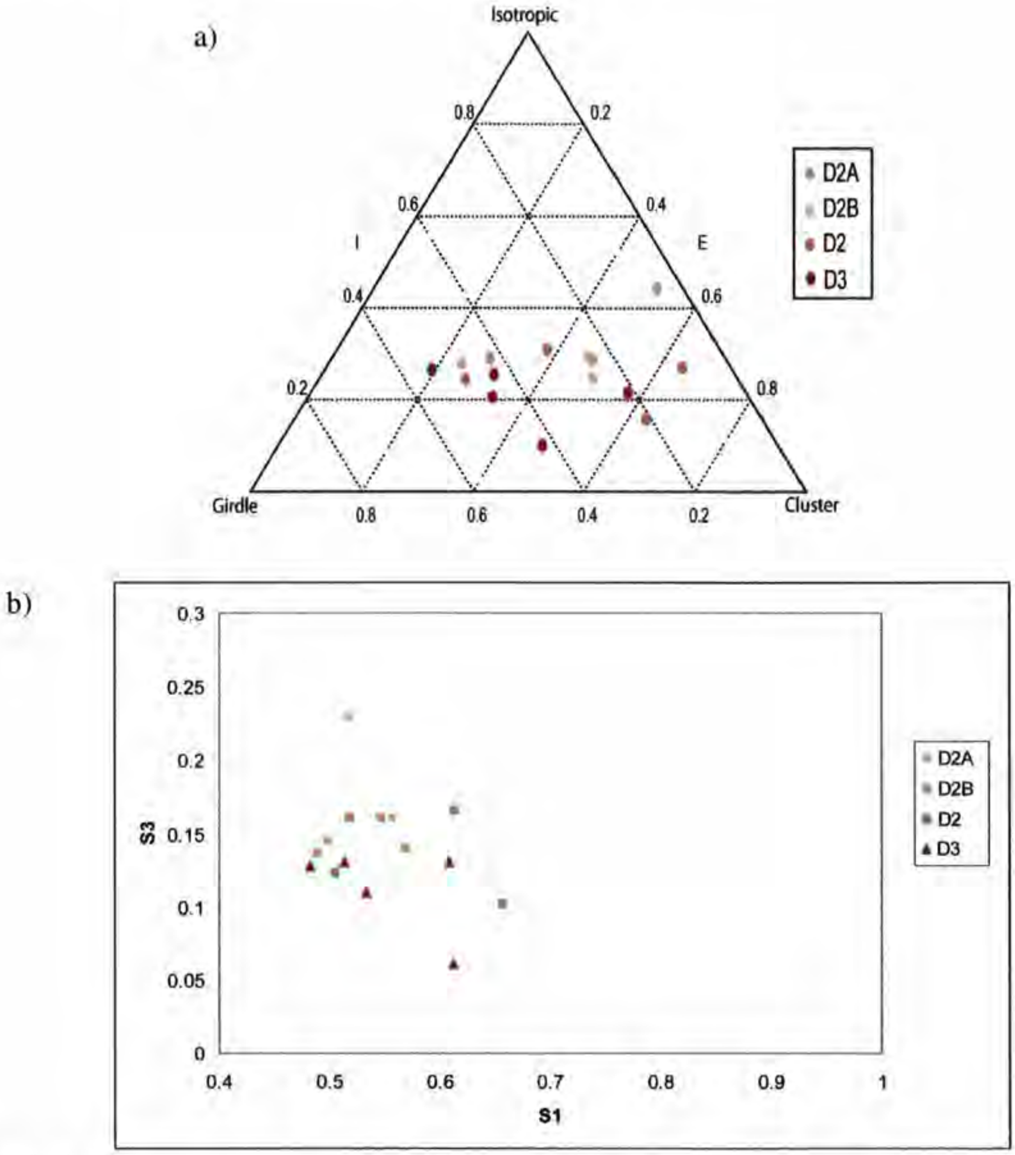
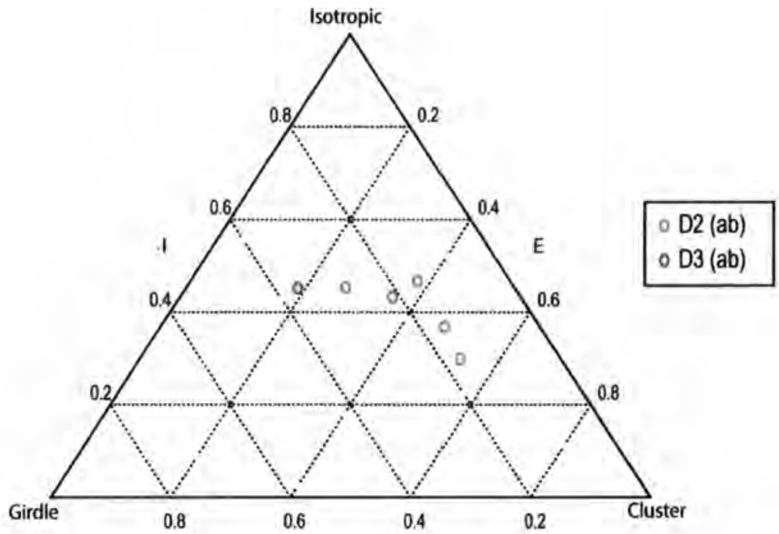


Figure 24: Eigenvalue plots derived from the a-axis clast fabric data. a) Ternary diagram and b) S1/S3 plot. Samples are colour coordinated according to the LFA they were sampled from.

The eigenvalues plots from the a-b plane data (Figure 25) plot higher on both graphs than the a-axis data. On the ternary diagram, the majority of points plot in the centre towards the right. On the S1/S3 plot the a-b plane data plot between 0.15 and 0.25 to the left of the graph showing considerably weaker fabric strength than the a-axis data.

a)



b)

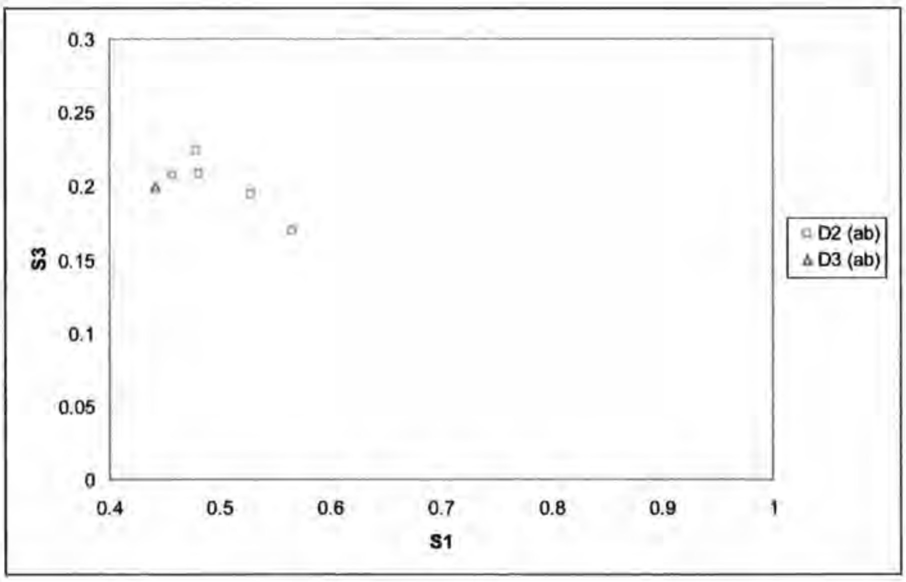


Figure 25: Eigenvalue plots derived from the a-b plane clast fabric data. a) Ternary diagram and b) S1/S3 plot. Samples are colour coordinated according to the LFA they were sampled from.

4.4 Clast Lithology

Clast lithology has been taken to establish the provenance of the diamicts. It has also been used to correlate and differentiate within and between diamicts. Sampling locations are marked on Figure 6. The percentage of each lithology within each facies can be seen in Table 3, with the expanded table in Appendix 1.

Table 3: Table showing the percentage of each lithology counted depending on each sedimentary facies.

Geology/Site	% D1	%D2	%D2A	%D2B	%D3	%SG
Limestone (Magnesian)		30.26	28.09	33.95	45.74	44.29
Shale	55.80	35.72	41.51	28.31	11.15	2.90
Sandstone (Quartz)		7.48	6.56	10.30	11.86	33.16
Mudrock	43.87	0.17	0.19	0.11		0.06
Sandstone (Lithic)		6.87	6.66	8.03	5.97	1.01
Limestone (Carboniferous)	0.33	3.41	3.47	3.80	7.39	2.43
Greywacke		4.53	4.05	4.99	3.34	2.66
Quartzite		2.71	1.83	2.28	1.99	3.08
Quartz		1.44	0.77	1.74	1.21	3.43
Siltstone		1.53	1.25	1.74	2.34	0.30
Wacke		1.47	1.93	1.19	0.64	0.18
Coal		0.46	0.29	0.87	2.20	0.41
Dolerite		0.72	0.39	0.98	2.06	1.78
Ironstone		0.75	0.39	0.33	0.92	0.53
Porphyry		0.43	0.29	0.22	0.64	1.84
Granite		0.26	0.10			0.12
Andesite		0.32	0.39		0.64	0.47
Old Red Sandstone		0.20	0.29		0.21	0.06
Rhyolite		0.29	0.19	0.22	0.50	0.30
Oosparite		0.23	0.58			
Sandstone (Arkose)		0.12	0.10	0.33	0.07	0.24
Flint		0.17	0.29	0.11	0.14	0.24
Diorite		0.09	0.10	0.11		
Chalk					0.14	
Pink Rhyolite		0.03			0.50	0.24
Breccia		0.03		0.11	0.07	
Porphritic Rhyolite		0.06	0.10			
Pyrite						0.06
Phosphate		0.12	0.19	0.22		0.06
Hematite		0.03				
Chert		0.03		0.11		0.12
Microgranite		0.06			0.28	0.06

4.4.1 Provenancing

Clasts found within the diamict are predominantly sourced locally, although in the upper diamicts and sands and gravels LFAs far travelled erratics have been found. D1 consists exclusively of local bedrock, shale and mudrock. This is part of the Jurassic sequence that makes up the solid geology in north east Yorkshire (Fox-Strangeways and Barrow 1915).

D2, D3 and the sands and gravels (SG) are made up of a wider range of lithologies although those that are locally sourced still dominate: shale and Magnesian limestone. Both till layers exhibit similar compositions of lithologies. The percentage of shale is marginally greater in D2A than D2B and significantly greater than D3. Both these upper tills and the sand and gravels layer have high proportions of Magnesian limestone and sandstones. The sand and gravels LFA has a higher proportion of sandstone than either of the tills. Other locally derived clasts within these layers are ironstone, siltstone, mudrock, oosparite, phosphate nodules, quartz and quartzite. These lithologies are all part of the bedrock Jurassic sequence (Kent 1980).

However, these LFAs also have many none locally sourced lithologies. It is the provenance of these clasts that will allow ice flow directions to be constructed. All three LFAs contain coal, carboniferous limestone and dolerite clasts which are derived from Northumberland and County Durham to the north of the field site (Johnston 1995). Flint from offshore is also found within the section from the Cretaceous beds of the North Sea (BGS Offshore map). The flint clasts have been found mainly within D2 and SG, with only one clast found within D3 which could have been derived from the LFAs below. The other main source of erratics is the Cheviot Hills. Porphyry, rhyolite, andesite and granite are all sourced from this area. While the last three lithologies mentioned can also be attributed to the Lake District, the pink nature of the rhyolite indicates the presence of feldspar, and is typical of these rocks in the Cheviots. The pink rhyolite is mainly found in D3, however, the darker red/purple rhyolite found in D2 is also indicative of the Cheviots (Robson 1976). While the black or dark grey nature of the andesite separates it apart from the dark green, purple and basaltic andesite of the Lake District (Robson 1976). Porphyry is only found in the Cheviots or south east Scotland. However, the granites found can be attributed to either the Cheviots or the Lake District. Both areas have descriptions of granites with concentrations of biotites and pink phenocrysts of feldspar as have been found in the samples from Upgang. Erratics that are distinctive to the Lake District are the microgranite, found in D3, and Diorite, which is found in very small quantities in D2 (Mosley 1978). While the main source of the greywacke found in the samples is the Southern Uplands in Scotland (Greig 1971).

Figure 26 shows a plot of the percentage of locally derived clasts against the percentage of those distally derived. This shows a negative correlation between the two classes upwards through the sequence. Samples from D1 plot in the bottom right hand corner, showing a very high percentage of locally derived clasts. Following this towards the left is the samples from D2A then D2B and D2, showing that the lower facies in D2 has the higher proportion of local material. Intermixed with these points is the sands and gravel LFA due to the high proportion of sandstone, which has been classified as a local lithology, within each sample. While this sandstone is a local lithology associated with the Jurassic sequence of northern Yorkshire, the local bedrock at Upgang is predominantly shale and mudrock. This indicates that the sandstone and, hence, the sands and gravels LFA, has been derived more distally. Samples from D3 show an increase of distally derived clasts compared to those sediments below, plotting towards the top left hand corner. The graph has been given an artificial regression of -1 as percentages have been used, the two values will always add up to 100. However, as each sample is of a different size, using percentages has allowed the change in distally compared to locally derived lithologies to be seen more clearly.

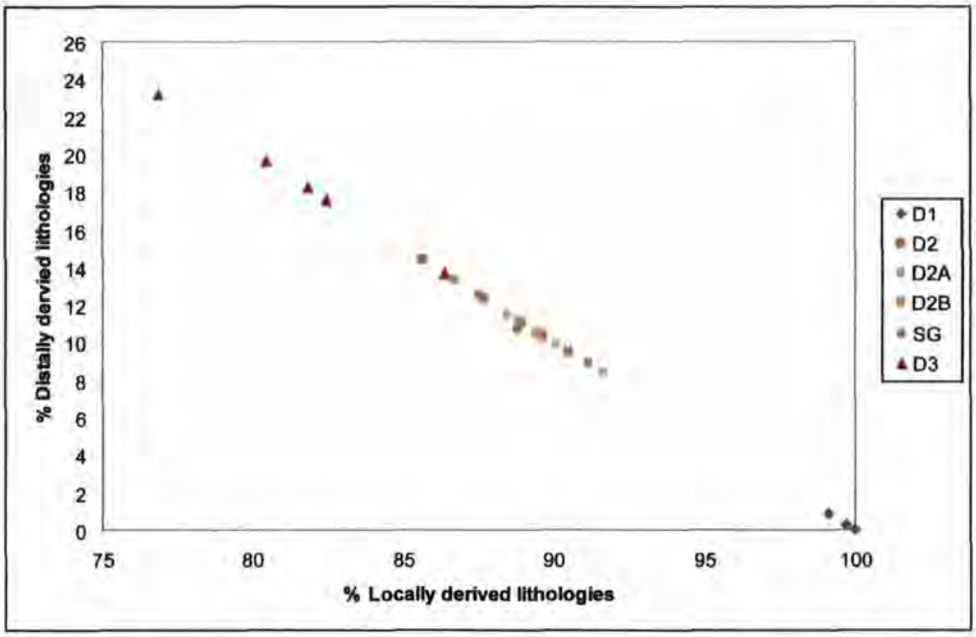


Figure 26: Scatter plot showing the percentage of local compared to far travelled lithologies in each sample.

4.4.2 Chi-squared testing

Chi-squared testing was applied to the data set in order to test correlations between samples in the same LFA and between LFAs. This is done by comparing each lithology with the mean number of clasts in each lithology class, thus indicating how each sample differs from the mean (Davies 2002). As discussed in Section 3.4 the closer the reduced chi-squared number is to 1 the stronger the correlation. The results obtained are shown in Table 4. Results have not been shown for the tests carried out using every lithology as these showed extremely low correlations. Due to the nature of the test it does not work well where a data set has a large number of rows or columns with zeros in them, as is the case with lithological samples which contain a range of erratics. A mean is calculated for each lithology, from all the sites being tested, meaning that at some sites the observed value differs greatly from the expected one. Thus, the whole sample is seen to differ largely from what is expected. However, in using only the locally derived lithologies, anything over 10 or 20% of the sample, chi-squared values can help to correlate and differentiate tills by showing how similar the quantities of each lithology is at each site. This clearly shows any differences between or within diamicts based on dominant lithologies, rather than focusing on changes in erratics of which only 4 or 5 clasts may appear in the whole section. Results from the percentage data appear to show stronger correlation than those from the absolute data. As each sample has a different number of clasts within it the expected values derived from the absolute data are skewed by the larger samples. For example, a sample of 400 clasts will have more shale clasts than a sample of 300. As a result the chi-squared value from the absolute data would show a poor correlation, yet both samples may have the same proportion of shale within them and are hence similar. Therefore, conclusions will be drawn using the percentage data as it is the proportions of each lithology that are important in correlation.

Table 4: Table of chi-squared valued comparing the clast lithological data over various facies and sites.

Test	X^2 (using %) (2 D.P.)	Degrees of Freedom	Reduced X^2	X^2 (using absolute values) (2.D.P.)	Degrees of Freedom	Reduced X^2
Lithologies ≥ 20% (limestone and shale)						
D2	36.13	9	4.01	116.12	9	12.90
D2A	2.61	2	1.31	9.08	2	4.54
D2B	9.6	2	4.80	28.75	2	14.38
D2A + D2B	117.05	5	23.41	445.74	5	89.15
D3	61.55	4	15.39	181.35	4	45.34
D3 (without 3.1)	1.63	3	0.54	4.26	3	1.42
SG	9.58	3	3.19	38.47	3	12.82
Lithologies ≥ 10% (limestone, shale and sandstone)						
D2	58.24	9	6.47	186.14	9	20.68
D2A	209.47	2	104.74	727.45	2	363.73
D2B	381.65	2	190.83	1421.25	2	710.63
D3	68.16	4	17.04	201.37	4	50.34
D3 (without 3.1)	3.48	3	1.16	9.26	3	3.09
SG	11.11	4	2.78	44.16	3	14.72
Comparison across lithofacies (Limestone and Shale)						
D2 vs. D3	175.22	13	13.48	516.92	13	39.76
D2A vs. D3	131.01	6	21.84	382.41	6	63.74
D2B vs. D3	86.6	6	14.43	249.56	6	41.59
D2A + D2B vs. D3	142.48	9	15.83	414.61	9	46.07
SG vs. D3	10.74	7	1.79	41.25	7	5.89
SG vs. D2	172.47	13	28.75	687.32	13	52.87
SG vs. D2A	108.34	6	18.06	423.66	6	70.61
SG vs. D2B	88.98	6	14.83	348.28	6	58.05
SG vs. D2A +D2B	141.28	9	23.55	559.54	9	62.17

Samples taken from within D2 (including those from D2A and D2B) show a fairly strong correlation between them. Samples from D2A and D2B can also lead to the classification of two separate facies on the basis of the quantity of shale and limestone within each sample. When tested as two separate facies a strong correlation pattern is seen, especially within D2A. In order to verify this, samples from both D2A and D2B have been compared, excluding those that have not been distinguished as A or B. The resulting chi-squared value is significantly large indicating that they are two different facies.

D3 correlates as one single till; however, the anomalously large proportion of shale in sample 3.1 obscures this correlation. Therefore, it has been excluded when correlating D3 to other till units. The four sand and gravel samples show correlation with each other suggesting they are of the same derivation.

When cross correlating between LFAs it can be shown that D2 and D3 are distinctly different from one another, something that can be seen in comparing shale values in the table. SG appears to be more closely correlated with D3 when looking at the locally derived proportion. However, as a sand and gravel facies it has different methods of eroding underlying bedrock to a glacial deposit. The far travelled component also needs to be looked at before it can be correlated confidently to either LFA.

It can easily be seen from Table 3 that the LFAs differ depending on the amount of shale present within them and, hence, the chi-squared correlations rely upon this value. This in itself is problematic as difficulties occur when counting shale and values can depend on which size classifications are used. Shale breaks up at almost every stage of the clast lithology identification; extraction, transport, washing and sieving. Therefore, the value can in some cases be anomalously high, although care is taken to try and minimise this.

4.5 Geochemistry

The geochemistry from Upgang has been analysed using chi-squared testing and cluster analysis. Sampling sites are marked on Figure 6.

Chi-squared testing has been used to correlate samples within and between LFAs. The test was carried out on all 43 elements and then on just the high and just the low abundance metals (Table 5). The results obtained when using all of the elements in calculations have very high reduced chi-squared values when comparing both samples within and between facies, showing a very weak correlation. When separated, the high abundance elements again show very high reduced chi-squared values, indicating very poor correlation. However, the results from the low abundance elements have values much closer to one and hence have a much stronger correlations. These results show a strong correlation within D2 and D3 along with a strong correlation between the two diamicts, inferring that they are geochemically very similar. D1 is geochemically distinct from both D2 and D3 emphasising the other differences between these LFAs.

Table 5: Table of chi-squared values from the geochemistry results. The closer the reduced x^2 number is to 1 the closer the correlation between samples.

	All		High		Low	
	x^2	Reduced x^2	x^2	Reduced x^2	x^2	Reduced x^2
D2	30861.21	771.53	30555.11	3395.01	19.93	0.64
D3	8516.09	212.90	8316.42	924.05	8.75	0.28
D2A vs. D2B	430.29	10.76	391.41	43.49	6.40	0.21
D1 vs. D2	50575.95	1264.40	49996.94	5555.22	281.44	9.08
D2 vs. D3	41114.02	1027.85	40604.67	4511.63	34.03	1.10
D1 vs. D3	28484.42	712.11	27929.10	3103.23	264.59	8.54

Cluster analysis on the whole geochemistry data set agrees with this similarity between D2 and D3 (Figure 27). Clustering between samples at the lowest level of the graph does not occur within the diamicts found during sediment description. Samples from D2 cluster with samples from D3 rather than other samples from D2 at this base level. This lower level clustering arguably forms a group including samples 2.1A, 2.1B, 2.5, 3.1 and 3.2 which could further be split into two. Sample 2.7 is fairly dissimilar from this main group of samples. Sample 1.1 is the most distinct sample with the highest dissimilarity measure mirroring the chi-squared results. This dissimilarity value represents the geometric distance between clusters. Therefore, the larger the distance the more dissimilar the samples are from each other.

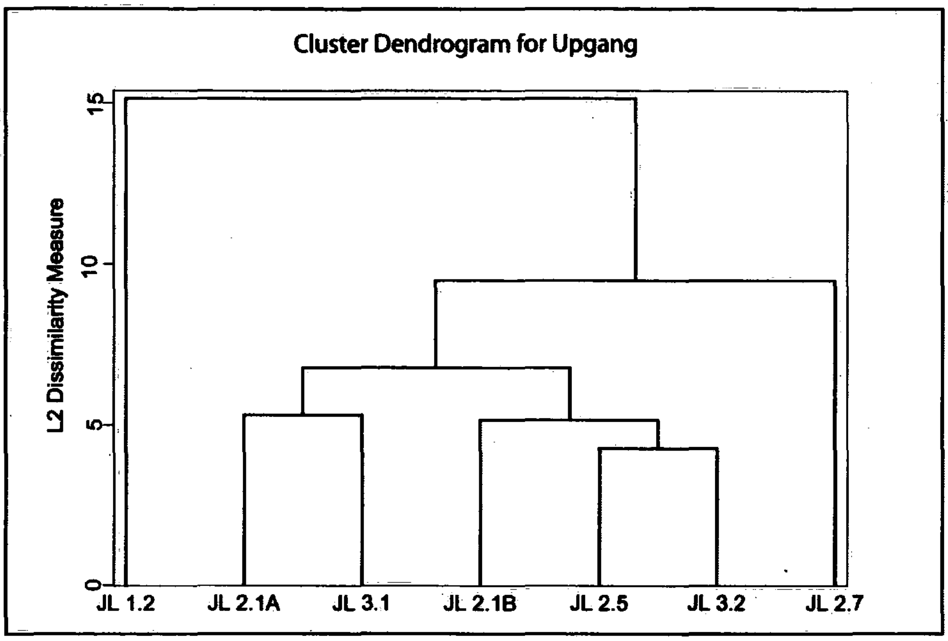


Figure 27: Cluster dendrogram constructed using the sediment geochemistry results. The closer the samples 'cluster' towards 0 the more similar the LFAs are according to geochemistry.

4.6 Geomorphological Map

Using the NEXT map DTM a geomorphological map highlighting moraines, glacial lakes, lineations and a potential drift limit in the North East of England has been created. Using this map in conjunction with the published BRITICE map it is possible to consider past ice margins and ice flow directions in the area. Features mapped from the BRITICE project are based upon published information mapping the location of the features in question (Clark *et al.* 2004). Locations mentioned within this section can be seen in Figure 28 (DTM of the area without any features marked on it).

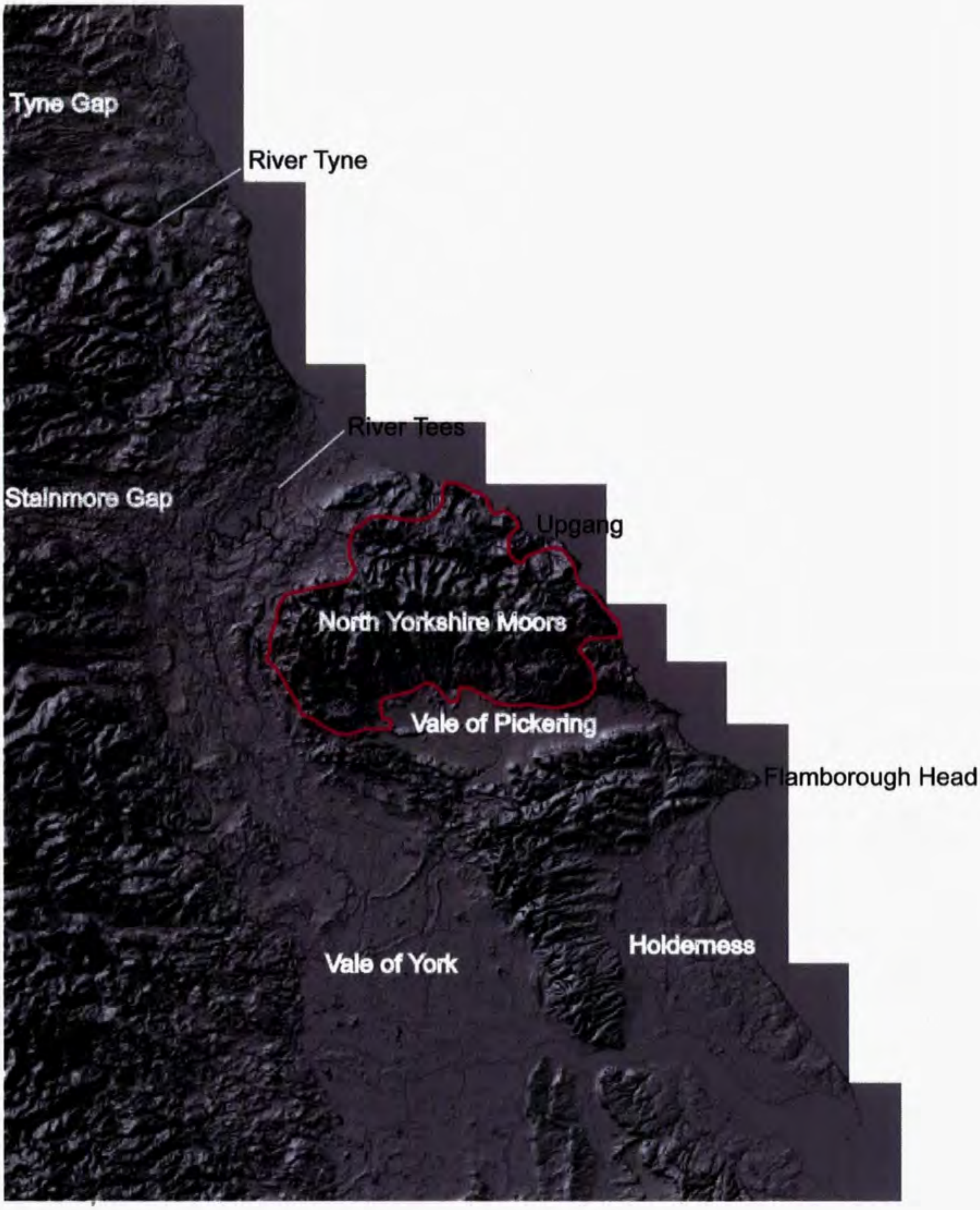


Figure 28: NEXT Map overview including sites mentioned within the text.

The map constructed showing the area around Upgang (Figure 29) shows very few geomorphological features. There is a potential linear feature to the east of the head of Sandsend Beck along with several eskers at the head of the Sandsend Beck. The existence of these eskers is confirmed by the BRITICE (Clark *et al.* 2004a) database (Figure 30). In addition, Figure 30 shows a series of glacial lakes in the highlands of the North Yorkshire Moors which are also described in Daysh (1958) based on sedimentological evidence. These eskers and lakes indicate the extent that the North Sea Ice Lobe came on shore with eskers representing the drainage system forming the lakes. However, from these maps it is unknown if the eskers are sub-, en- or supra-glacial. It is possible that this ice was forced onshore through the three river valleys seen along the coast (Figure 29). In comparing these valleys to those within the unglaciated North Yorkshire Moors they appear very different as though draped in glacial sediment. This sediment has not been considerably eroded by the rivers during the Holocene resulting in the softer, rounder nature of the watersheds. Therefore, these valleys can be seen as a pre-glacial feature in which sediment has been deposited as the ice came onshore. Aside from these features there is potentially some hummocky deposits to the north-west of Upgang, although there does not appear to be any directional features between the North York Moors and the coast that would indicate ice flow direction. When combined with the BRITICE map (Figure 30) there is still very little geomorphological evidence of ice existing in the area.

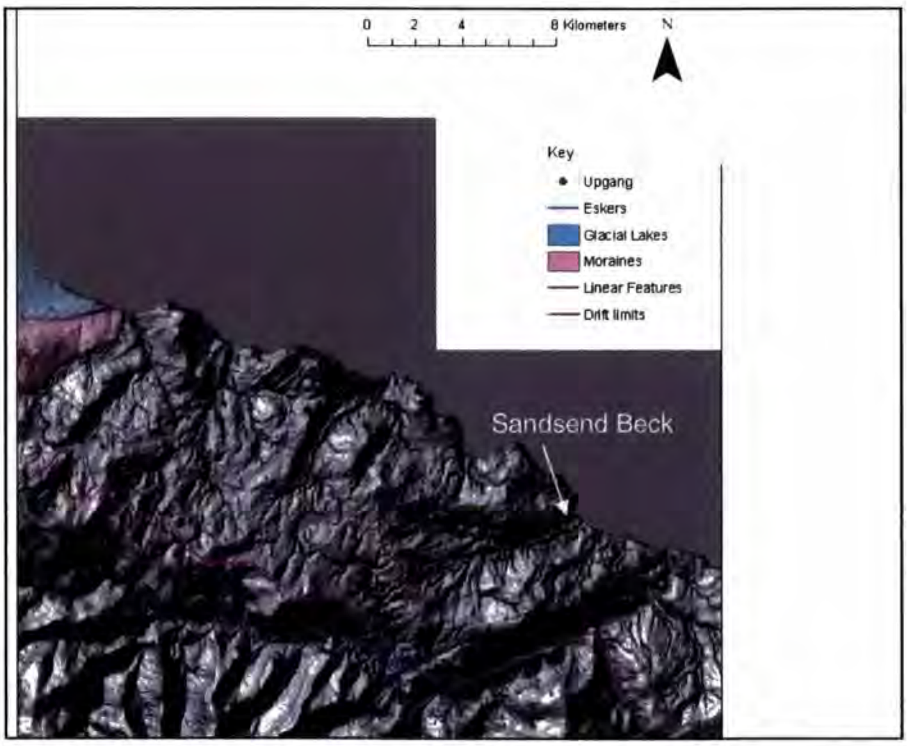


Figure 29: NEXT Map DTM including mapped features focusing on the Upgang area.

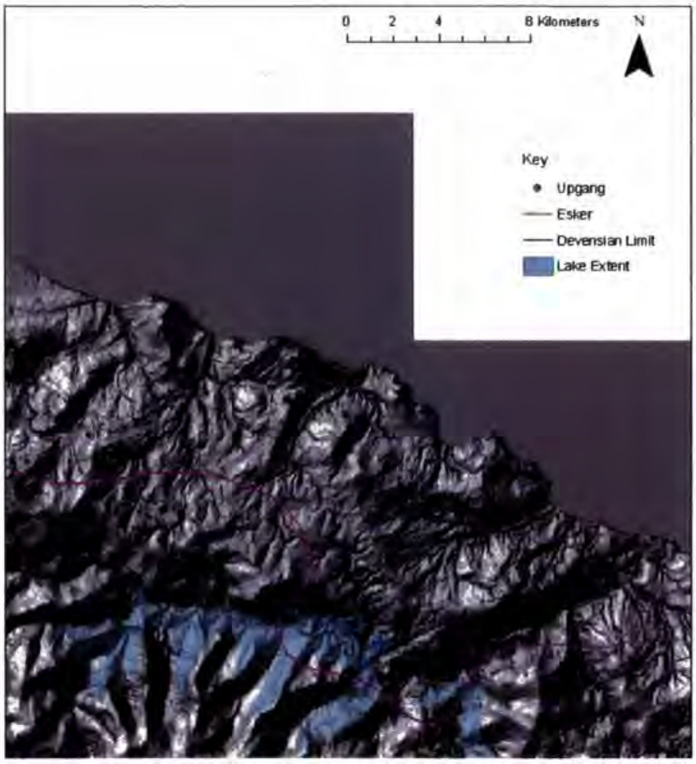


Figure 30: Map focusing on the Upgang area including features from BRITICE (Clark *et al.* 2004a).

Along with these features a drift limit has also been marked on the map (Figure 29). This has been constructed on the basis of a difference of shading seen on the NEXT Map DTM. Areas covered in sediment appear lighter with a smoother less angular topography. This drift limit comes on shore just south of Upgang and follows the lower edge of the highland that is incorporated within the North York Moors. Figure 30 shows the Devensian ice limit that has been constructed on the basis of the extent of the Skipsea Till and its equivalents (Clark *et al.* 2004). This limit is roughly similar to the drift limit in Figure 29.

Figure 31 shows the continuation of this ice limit throughout the North East of England as depicted on the BRITICE database. This limit has been defined using the onshore extent of the Skipsea till (Clark *et al.* 2004a). Starting in the west of the map the limit can be seen amongst the highland of the Pennines. The marked limit is then broken by Glacial Lake Humber before following the eastern edge of the Vale of York. As discussed above the limit follows the edge of the North York Moors and the highland area to Flamborough Head. It then pushes on shore to the raised area west of low lying Holderness. This ice limit has been supported by the identification of a drift limit on the NEXT map image around the north and eastern margin of the North Yorkshire Moors (Figure 32). This limit, which has been established on the basis of differences in shading and topography is very similar to the limit defined using BRITICE.

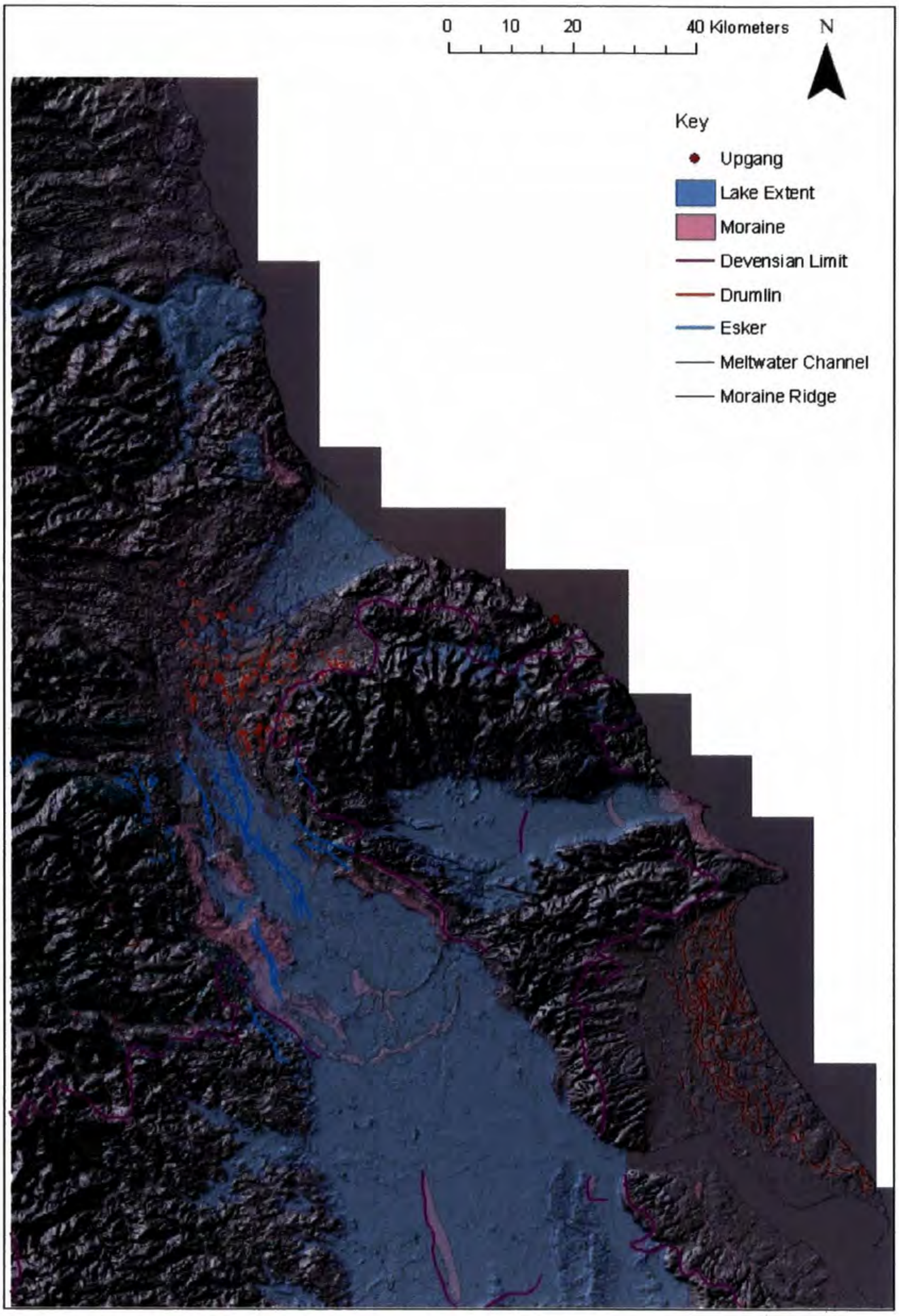


Figure 31: Map of North East England including the mapped BRITICE features from Clark *et al.* (2004a).

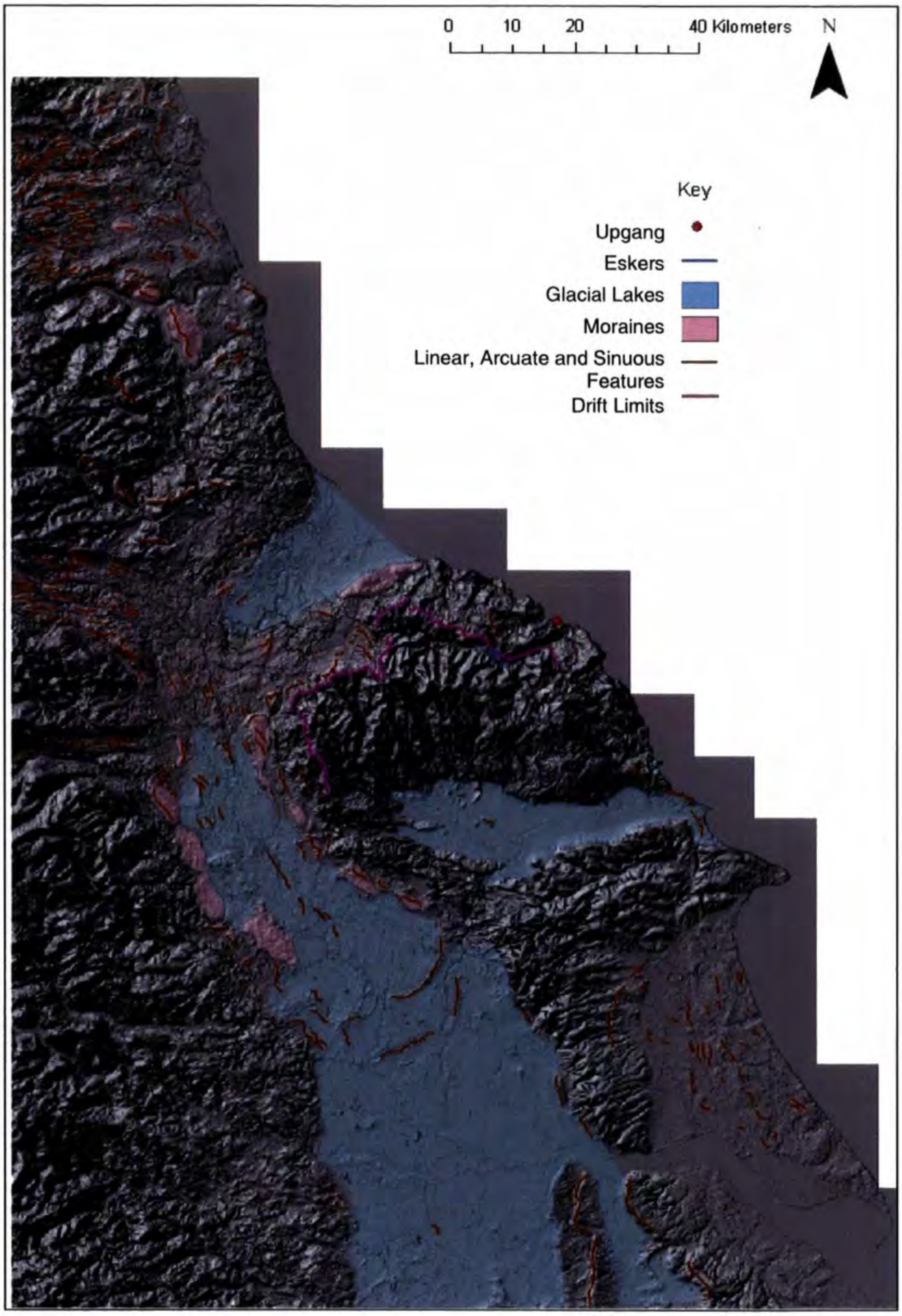


Figure 32: NEXT Map DTM of North East England including geomorphological features.

In addition to the LGM drift limit a number of other glacial features have been highlighted from the NEXT map DTM of the North East of England (Figure 32). Moraine features, shown in pink, can be seen along the northern edges of the Vale of York, both east and west marking the edge of a glacial lake. A moraine can also be seen to the south of Middlesbrough marking the edge of another glacial lake and a number of moraines can be seen to the north. Two of these moraines, north of the Tyne, are relatively small, while the moraine to the south of the Tyne is aligned north to south and considerably larger. The BRITICE data in Figure 31 supports the positioning of these moraine features along with the addition of other moraines at Flamborough Head and within the Vale of York.

Linear, arcuate and sinuous features have been marked on the NEXT map image throughout the map (Figure 32). A high concentration of these features, potentially lineations or drumlins due to their linear nature, aligned west to east can be seen in the Tyne and Stainmore Gaps. To the east of these gaps the alignment of the lineations alters to a north-west south-east direction. However, these lineations in the Tyne and Stainmore Gaps have not been depicted on the BRITICE map although an area of drumlinisation to the north of the Vale of York has (Figure 31). A number of these linear features can also be seen in the Vale of York aligned broadly north to south, (Figure 32), which appear to be eskers when compared to the BRITICE map (Figure 31). Another set of these linear features marked on Figure 35 towards the centre of the Vale of York appear to mark an ice limit lying almost perpendicular to this north-south direction suggesting that they are end moraines. This is supported by the BRITICE database (Figure 31) and their arcuate nature. Lineations have also been depicted lying north to south in the Holderness area in both Figure 31 and Figure 32.

In addition, to these features, several glacial lakes have been depicted in the area. Figure 32, shows three glacial lakes, Tees, Humber and Pickering identified as a result of the low lying terrain. This was supported by the BRITICE data (Figure 31) to confirm the existence of lakes in these locations. Lake Pickering is situated in the Vale of Pickering and is delimited by the highland on three sides of it, while Lake Tees, around the River Tees, only has one limiting moraine to the south suggesting it was an ice dammed lake. The largest lake, Glacial Lake Humber is limited by the highlands on either side of the Vale of York along with several lateral moraines. In addition, BRITICE depicts Glacial Lake Wear and a number of glacial lakes within the highlands of the North Yorkshire Moors. A series of meltwater channels on the western edge of the Vale of York have also been depicted (Figure 31).

5. Reconstructing the glacial history of Upgang

5.1 Diamict 1 - genesis and provenance

D1 is a grey brecciated bedrock lithofacies that can be seen at the base of the sequence at Upgang. As discussed in the previous chapter the lithology of this facies consists of Jurassic shale and mudstone from the bedrock that underlies the Quaternary sediments around Upgang. The clast form data from this unit shows that the clasts are blocky with high RA values. Subsequently they plot in the top left hand corner of the covariance plot (Figure 33). High RA values suggest that they were transported only a short distance and low C40 values indicate that they have been subject to subglacial processes. When compared to the covariance plot produced by Benn (1992) the graph from Upgang shows D1 plotting in a similar area to bedrock and raft fragments. However, the results from Benn (1992) are from clasts of a crystalline lithology and hence have been altered during transport differently to those at Upgang. Therefore, a hypothesis suggesting that D1 is a facies of local bedrock fragments can be formed due to the provenance and lithology of the unit. The facies is also geochemically distinct from the overlying sediments highlighting the local nature of the sediments.

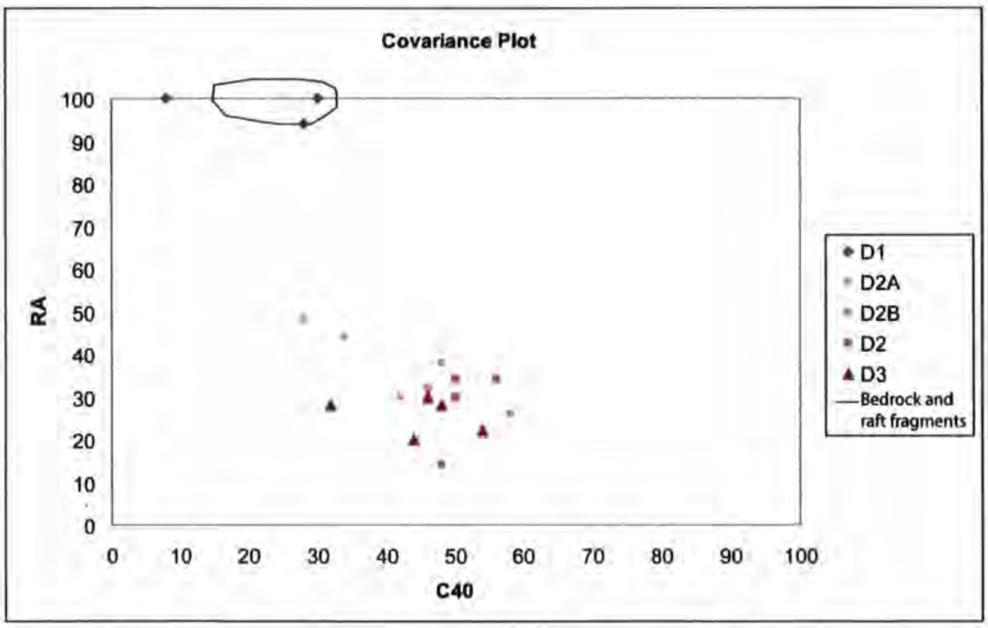


Figure 33: Covariance plot from clast form data at Upgang. Envelope adapted from Benn (1992).

However, the genesis of the facies is less clear. It could be described as a glaciotectonite formed by subglacial processes. Whilst it is also possible that the brecciation of the bedrock was emplaced initially as a result of proglacial thrusting or that the LFA was deposited during a glaciation prior to the Last Glacial Maximum (LGM).

A glaciotectonite is defined as:

> 'Rock or sediment that has been deformed by subglacial shearing (deformation) but retains some of the structural characteristics of the parent material.' (Evans *et al.* 2006 p 169).

As a result D1 can be classified as a glaciotectonite due to the barely broken up bedrock nature of the facies. The weak sedimentary Jurassic rocks have been subglacially detached from their source and entrained resulting in the destruction of primary structures and brecciation before deposition (Elson 1985). In some places along the section the rock is very highly brecciated, while in other areas the facies consists of solid shale and mudrock rafts. Due to the softness it is likely that it would have only been transported a short distance before the rock was deposited and set, which is supported by the high RA values. Pederson (1985) described the processes that form a glaciotectonite in a three step model. Firstly jointing and fracture occur, which is followed by the development of shear planes, and finally the continuous crushing and grinding of fragments to create a fine grained matrix. Lithology and pore water pressure play an important role in this process assisting in deformation. High pore water pressures can result in deformation at low strain rates (Evans *et al.* 2006).

It is possible that these processes occurred as the glacier deposited D2 and an analogy could be drawn with similar material at Clougher Bay (Hiemstra *et al.* 2007). However, at Clougher Bay the glaciotectonite is separated from the diamict by a layer of heavily sheared bedrock. This communition till layer (Benn and Evans 1996) is not present between D1 and D2 at Upgang. One proposal to explain this rapid vertical transition in deformation styles between glaciotectonite and overlying diamict are of 'jumps' in subglacial strain rates representing a rapid switch between dialant and non-dialant conditions within the till (Alley 1989). An increase in strain rate results in dilation and the weakening of sediments. Yet as the strain rates then fall the sediment collapses and strengthens, leading to even lower strain rates causing a positive feedback mechanism. It is these feedbacks that Benn and Evans (1996) used to explain the sharp vertical contacts between glaciotectonites and deformation tills. However, due to the recorded occurrence of communition till, for example at Clougher Head (Hiemstra *et al.* 2007), these 'jumps' are thought to depend upon particular subglacial conditions or sediment granulometry (Benn and Evans 1996). The nature of D1 would limit the amount of water from the overlying ice that could be drained through it allowing basal water pressures to build up. The presence of these high basal water pressures would lead to glacier decoupling preventing the transmission of stresses within D2 down into D1 (Evans *et al.* 2006). It is therefore these differences in lithology between D1 and D2 that are partly responsible for the formation of the décollement plane (Figure 12).

Such distinct décollement planes can also represent a sedimentary boundary between pre- or pro-glacial sediments and sub-glacial tills (Roberts and Hart 2005). As a result, D1 could have been proglacially thrust into this position. Due to the nature of the sediment it is possible that brittle deformation took place during the proglacial thrusting (Hart and Boulton 1991). A décollement boundary could then have been formed as the ice depositing D2 overran the area resulting in the sharp planar boundary between D1 and D2. The production of this décollement plane could have been a result of very high water pressures, low strain rates or a decrease in basal shear stress (Alley 1991). Piotrowski *et al.* (2001) suggest that these sharp contacts between a till, for example D2 and the underlying substratum, D1, are very common in the glacial record. While it is possible that D1 was deformed proglacially by thrusting and stacking, the deformation could also have occurred subglacially after initial proglacial deposition of the shale facies as ice initially overran the area. This would require similar mechanisms to those described above existing for the formation of a décollement boundary to occur. Catt (2007) suggested that as the bedrock lithology immediately inland of Upgang is Jurassic sandstone it is likely that D1 was derived from the sea floor to the north where the lithology is dominated by shale. This would appear to support a proglacial origin to the LFA. However, Catt (2007) suggested that due to the provenance of the LFA it was likely to be a till, supporting the glaciotectonic hypothesis.

An alternative explanation for the existence of D1 is provided by Hemmingway and Riddler (1980). Two liassic shale rafts existing within the same till horizon were described as similar to D1 from this report. One raft was 98 metres long and 3.2 metres thick while the other was 30 metres long with an unseen base. As the base of D1 was never found during this investigation it is possible that the facies is part of the series of rafts described by Hemmingway and Riddler (1980). These rafts were thought to have been transported subglacially from the submarine Cleveland Ironstone Formation, although no time frame for this event was proposed. It was suggested that the ice lobe was confined by the local topography favouring the shearing of blocks from this escarpment. The rafts were then driven upward and landwards to their position in the cliff. However, as the base of D1 was not uncovered during this investigation along with the facies outcropping for considerably more than 128 metres this proposed genesis cannot be confirmed.

Another explanation stemming from the deposition of D1 is that it may not have been deposited during the last glacial period. Evidence for pre-Devensian sediments are found elsewhere along the north east coast. For example, the Basement Till at Dimlington is believed to have been deposited prior to the last glacial as it underlies the Dimlington silts, dated to 21 000 cal. years BP (Madgett and Catt 1978). A considerable amount of time is believed to have passed between the deposition of the Basement Till and the overlying Skipsea Till, including the Ipswichian interglacial (Catt and Penny 1966). Therefore, D1 could have also been deposited by a pre-Devensian glaciation as favoured by Catt (2007). This was proposed due to what was considered to be a weathering profile in the upper 40-60 cm of the diamict, described as being a paler colour to the lower section and weakly oxidised (Catt 2007). Despite this proposal, within this study D1 appeared to be the same colour through-out the horizontal and vertical profile. However, direct dating control of the site and the sediments would be needed in order to pursue this model further. This model is weakened by the amino acid dates presented which date the Basement Till to c. 20 000 years BP (Eyles *et al.* 1994). Hence, a Devensian overrun proglacial thrust feature or a glaciotectonite genesis for D1 is the most plausible explanation.

5.2 Diamict 2 – genesis and provenance

As described in Section 4.1 D2 is a partially stratified brown sediment found in the lower section of the cliffs with the stratification denoted by a crude change in colour. The sediment logs and photographs along with the quantitative data suggest that this is a subglacial deposit, deposited as part of a deforming bed.

The clast form data from this facies plot around the 0.4 line of the ternary diagrams (Figure 17), suggesting that the clasts have been altered subglacially. This is supported by the covariance plot that shows samples from D2 plotting in the area of the graph where morainic deposits are considered to dominate (Benn and Ballantyne 1994), (Figure 34). This suggests that not all sediment was transported subglacially and that it could include some supraglacially or englacially transported sediment. However, as with the graph from Benn (1992) used to highlight the raft nature of D1, the lithologies samples in Benn and Ballantyne are predominantly crystalline and, hence, have been affected differently to the sedimentary rock lithologies predominantly found at Upgang as a result of transport processes. Other lines of evidence discussed within this section will clarify the origin of this diamict more clearly. Samples taken from the lower facies of D2A plot higher on this covariance plot than those from D2B which indicates that the samples from D2A have a high RA value. This suggests that the clasts in D2A are blockier than those in D2B, which could indicate the greater inclusion of local bedrock as the samples from D1 are blocky clasts. While the clast form data indicates a possible subglacial origin for the diamict a clearer depositional history cannot be established using this method as deformation, lodgement and meltout tills all have similar clast forms (Bennett *et al.* 1997). Therefore, the clast fabric data and various features from the sediment logging need to be considered before any conclusions can be drawn.

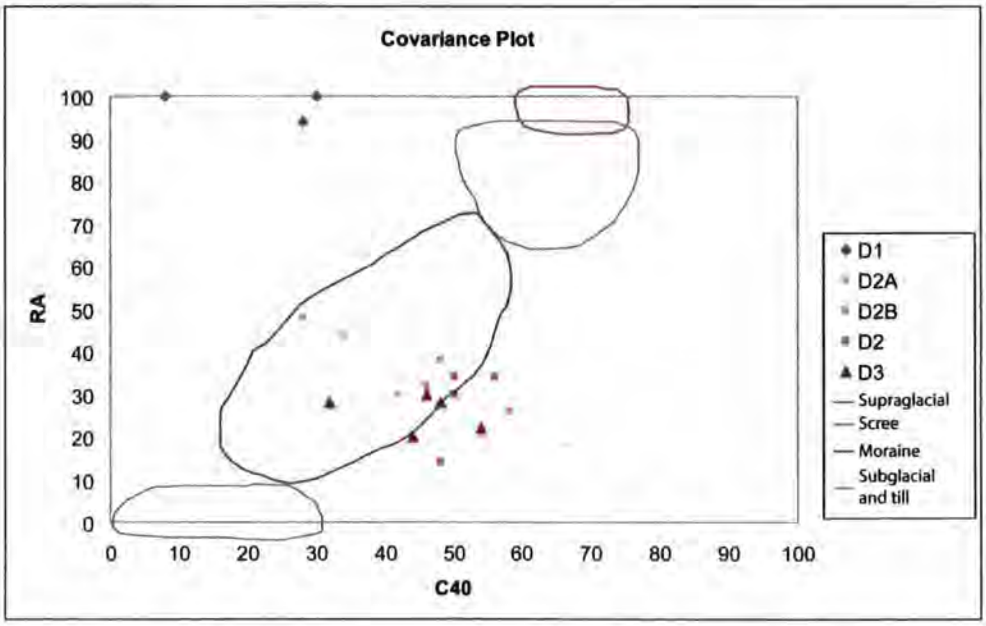


Figure 34: Covariance plot from Upgang. Envelopes adapted from Benn and Ballantyne (1994).

The clast fabric data supports the hypothesis of a subglacial origin for the till. A modality-isotropy diagram (Hicock *et al.* 1996) has been constructed using the step contoured stereonets (Figure 35). The polymodal and spread bimodal clusters indicate that the sediments are subglacial tills formed by deformation, while the remaining points fall within the lodgement sector of the graph. As Evans *et al.* (2006) state that lodgement tills are a hybrid of deformation, ploughing and lodgement; these samples can therefore be said to show an increased lodgement component. However, there is a subjective nature to this method as often the difference between clustered bimodal and spread bimodal can be subtle. This is supported by the polymodal nature of the a-b plane data and suggests that the few samples with increased lodgement components could be anomalies, especially when they are located next to the a-b plane data.

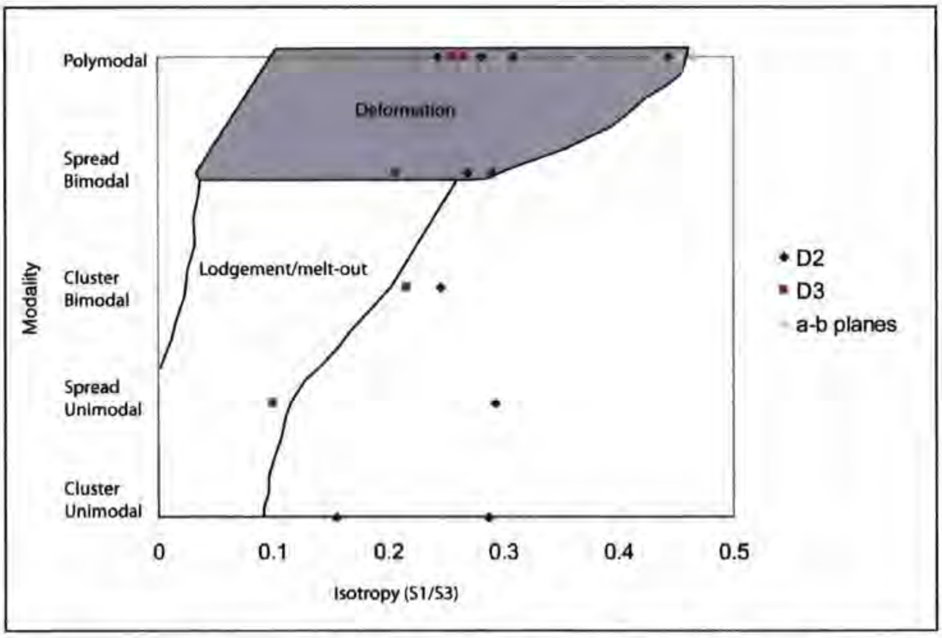


Figure 35: Modality-Isotropy Diagram for the samples taken at Upgang based on Hicock *et al.* (1996). Deformation and Lodgement-meltout envelopes from Evans (2000).

The ternary and S1/S3 plots (Figures 36 and 37) also indicate a deforming bed origin to the sediment. On the ternary plot, samples plot in the area Benn (1994) inferred as relating to non uniform deformation tills, the upper till studies at Breidamerkurjokull, Iceland. The relatively weak nature of the clast fabrics at Upgang suggest Type A deformation occurred as was observed at Breidamerkurjokull (Benn 1994). This is interpreted as ductile formation and is found at other places along the Yorkshire coast, for example Skipsea

(Benn and Evans 1996). The a-b plane data plots within the envelope marked as subglacial till by Evans *et al.* (2007). On the S1/S3 graph the samples plot in the area of the graph Dowdeswell and Sharp (1986) describe as deformed lodgement till. While on the S1/S3 plot Hart and Roberts (1994) suggest that the area of this graph in which the samples plot indicates homogenous soft bed till. By considering the Gaussian weighted stereonets a low variable dip can be seen for the clasts, which is a sign of a deforming bed unit. If the dip had been stronger in an inconsistent direction then a case could be made for a mass flow origin to the sediments. However, other factors should be considered besides clast fabrics as they can show great variation both between and within units (Bennett *et al.* 1999).

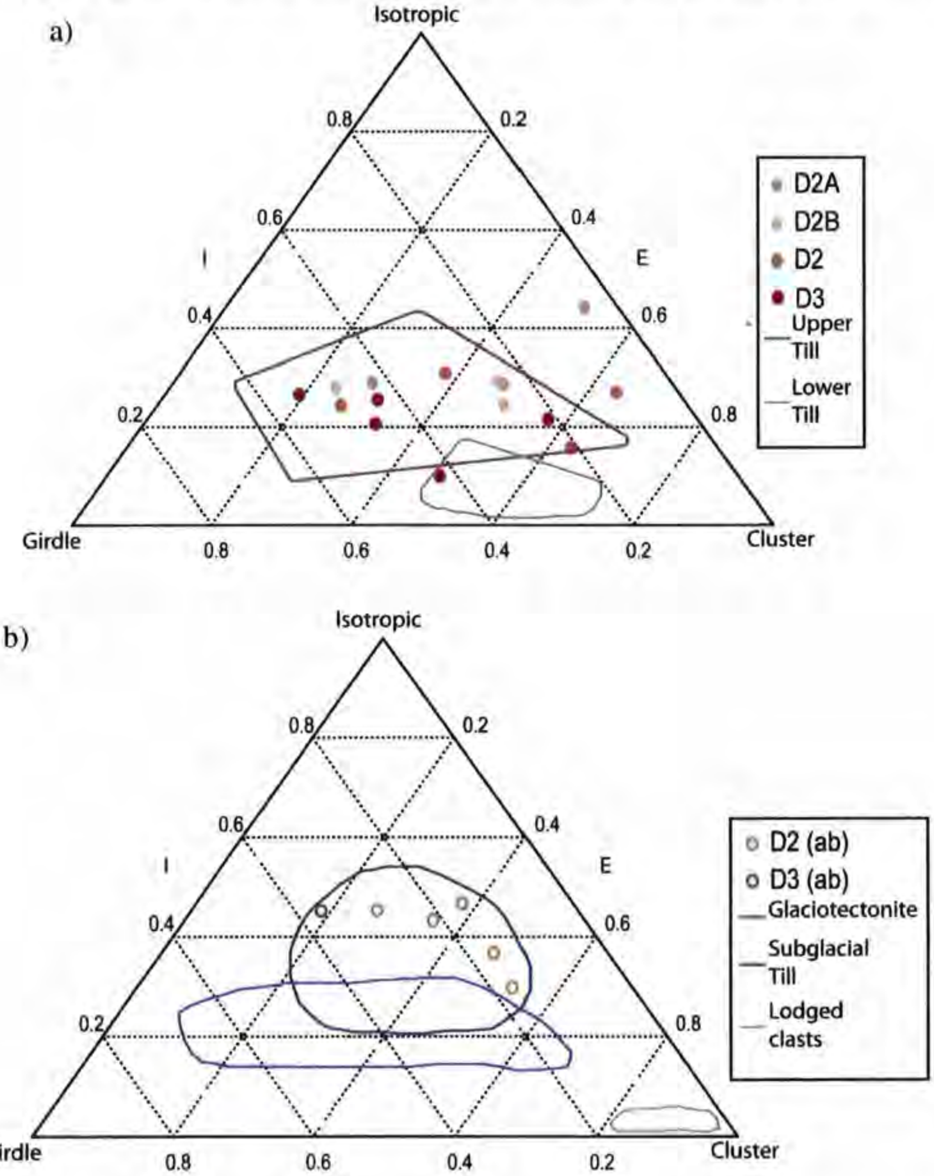


Figure 36: Ternary diagrams from a) a-axis data including envelopes from Benn (1994) and b) a-b plane data including envelopes from Evans *et al.* (2007).

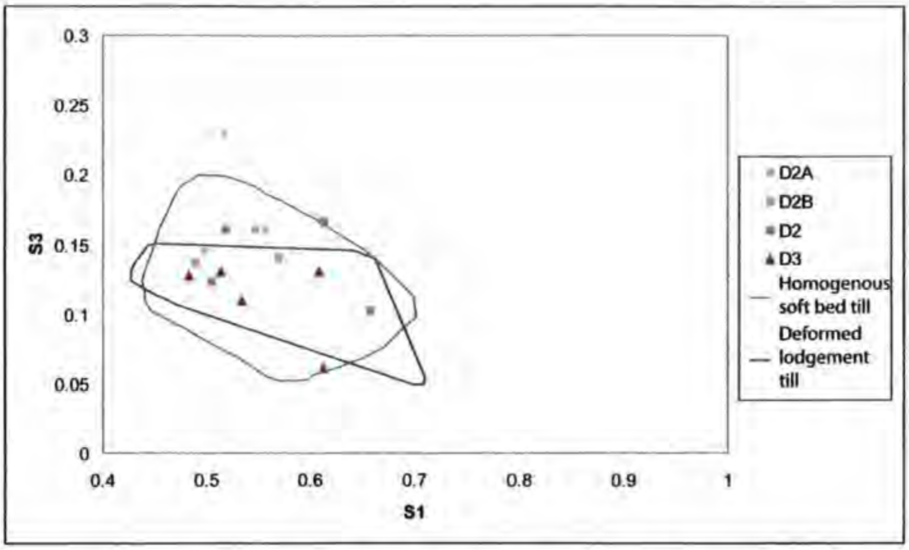


Figure 37: S1/S3 plot from a-axis data. Envelopes from Dowdeswell and Sharp (1986) and Hart and Roberts (1994).

Other features from the sediment logs can also be used to support this subglacial deforming bed theory. Firstly, the sand feature seen in Section H (Figure 38) can be interpreted as an inclusion that has been entrained within the sediment and subsequently folded and boudinaged as a result of shear stresses within the till. This feature can be compared to many found at the type site of the deforming bed in Britain at West Runton, north Norfolk (Hart 1994, Hart and Roberts 1994, Hart 1995, Roberts and Hart 2005). The clast fabric data taken from sample 2.5, which is located above the feature, shows a high fabric strength, which supports the evidence that this feature is a result of high shear stresses (Figure 38). Stress shadows should be found at either end of the boudin resulting in weak clast fabrics; strong fabrics should be observed at the sides as sediment is pushed around the feature thus increasing stress (Hart 1994). As the till has thickened during deposition the deforming layer will move up the sequence preserving the feature (Larsen *et al.* 2004). This can also explain the absence of any other boundinage features. It is possible that other features are not visible due to slumping of the upper sediments down the cliff covering them or that they have not yet been exposed. Although high stresses within the deforming layer would lead to the homogenisation of the till resulting in few or no features such as boudins (Alley 1991). One other similar feature is the fold seen in Section K (Figure 16). This isoclinal fold would have been deposited within the deforming bed as a result of simple shearing within the subglacial mass (Hart and Bolton 1991). It is likely that the darker diamict of D2A was detached from its bedding and then folded over by the shear stresses within the diamict at the time of deformation. The form of an individual feature is dependent on the strength of the sediment which forms it. For example, fine grained units can be highly and often complicatedly deformed, as seen in Section H, whereas more coarse grained sediments form relatively undeformed features, Section F (Hart and Roberts 1994, Benn and Evans 1998). These differences in style of deformation reflect the strain response of the various sediments to stress due to contrasts in frictional strength and permeability (Benn and Evans 1998). The fine grained sediments of Section H had a reduced frictional strength and high permeability resulting in the high level of deformation. The isoclinal fold in Section K suggests that the diamict that was folded into D2 also had low strength possibly due to a high permeability or low friction between the unsorted sediments. However, Piotrowski *et al.* (2001) suggest that the preservation of such sediments, which remain undisturbed internally with sharp boundaries, is not possible with a deforming bed model due to the expected diffuse mixing. Piotrowski *et al.* (2001) argued that due to the simple shear within a deforming bed features should loose their coherent appearance within a short time scale as a result of particle diffusion.

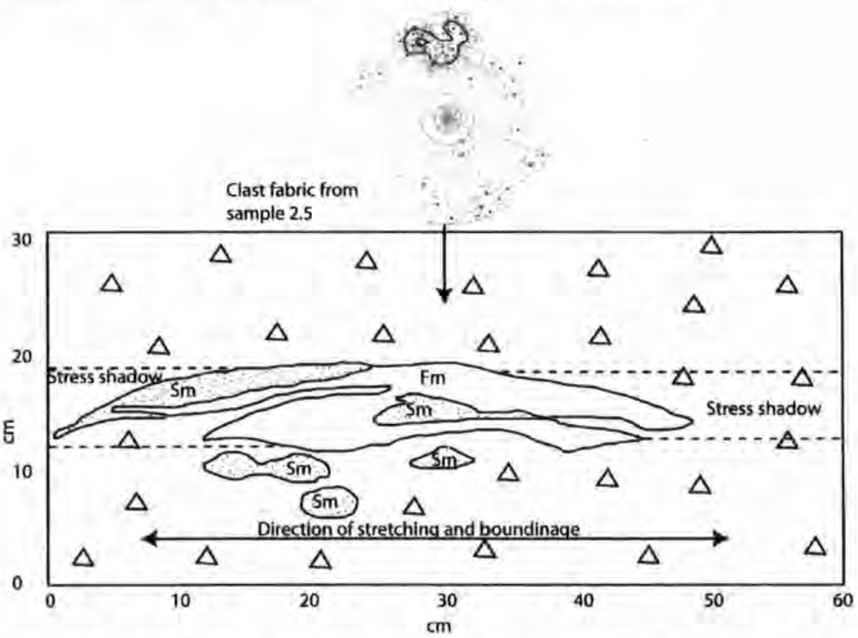


Figure 38: Annotated sketch of Section H showing clast fabric from sample 2.5 along with direction of boundinage.

Secondly, the crudely stratified diamict at the base of the sequence can also be classified as a deforming bed feature. Van der Meer *et al.* (2003) used this apparent development of banding in the till as an indication of a deforming bed. Hart and Roberts (1994) and

Roberts and Hart (2005) also described laminated sediments as indicators of tectonic deformation within a deforming bed. Under this model the sediment has been folded and subsequently exposed to high lateral strain causing them to 'stretch' out to form laminations (Figure 39). As the banding is thin and discontinuous a subglacial glaciotectonic model can be applied to D2. The boudin feature in Section H lies directly above this layer, supporting the levels of high strain, along with the clast fabric diagrams from sample sites 2.6 and 2.7 which also show high strain rates. As the till then grades into a massive unit it is suggested that the thickness of the deforming bed altered possibly as a result of changes in porewater pressure or the composition of the till matrix. Different matrixes, sand, silt or clay, result in different depths of deformation (Larson *et al.* 2004), however, the clast size of the matrix has not been analysed as part of this project. At this particular location it is difficult to find definitive evidence to show that this feature is a result of subglacial deforming bed conditions, although evidence within the same lateral unit (Section H and Section K) show tectonic deformation. Due to the ambiguous nature of these stratifications an alternative genesis could be deposition as a result of mass flow units (Zielinski and van Loon, 1996; Roberts and Hart, 2005). This process would lead to the stratified sediments as each bed would belong to a different 'flow' event. Benn (1996) interpreted the stratified diamict, underlying massive diamict, at Achnasheen Scotland as the result of subaqueous mass flows. It was suggested that the unit was saturated till from beneath the glacier front which then flowed beyond the margin as cohesive to cohesionless debris flows due to gravity. The flows were considered to be non-erosive and unchannelised due to undisturbed contacts with the underlying sediments. This description is similar to the stratification found in D2C. Therefore, this hypothesis could explain the genesis of this facies and why the stratified diamict is only found in a limited area at the base of the section. Despite this proposal very little other evidence taken from Upgang, supports a mass flow hypothesis suggesting that the stratification formed as a result of high shear strain and not as mass flow units.

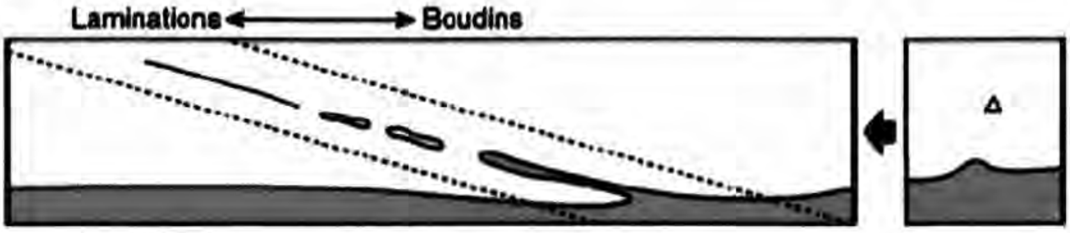


Figure 39: Subglacial glaciotectonic deformation of a large perturbation into folds pods and boudins after exposure to high shear strain. From Hart and Roberts (1994).

Thirdly, a raft of material from D1 has been shown in Section F (Figure 13). According to van der Meer *et al.* (2003) till rafts are another indication of a deforming bed. Ruszczynska-Szenajch (1987) suggested that these types of rafts can be formed by glaciotectonic processes that involve the rip up and deposition of the feature without being incorporated into the ice body. It is suggested that this feature could be a result of squeezing up, in which the raft was detached through décollement, transported a short distance, maintaining its primary structures, and deposited above the parent sediment unit during the squeezing up of the sediment along a shear plane. It is also possible that the raft was deposited as part of the deforming bed. As D1 was overrun it could have been ripped up, entrained and incorporated into the deforming bed before deposition. The shear stresses responsible for the boundinage of the feature in Figures 14 and 38 could have also caused the attenuation and boundinage of the raft in Section F. An alternative genesis for raft features is as 'products of englacial transport and deposition later from stagnant ice' (Piotrowski *et al.*, 2001 p 142). However, the similarity of the diamict within this feature with D1, along with its attenuated shape favours the rip up proposal for the genesis of this feature. Other considerably larger raft features were presented in Hemmingway and Riddler (1980) as products of subglacial shearing.

Another feature that is analogous to a deforming bed origin for the till is the sharp boundary between D1 and D2 (Figure 40). This sharp boundary can be described as a décollement plane in which D2 has overrun D1 without deforming or including the sediment within it (Alley, 1991; Hart and Roberts, 1994; Hart, 1995). It has been suggested that a low slope angle of the ice leads to the deformation of sediments as well as producing a décollement plane between the sediment deposited and the underlying unit. Alley (1991) listed several lines of evidence to support the correlation of a sharp basal contact and the deforming bed model, discussed in section 5.1. Alternatively, Piotrowski and Kraus (1997) suggest that a thin film of water separates subglacial sediments from the shear stresses imposed on them resulting in sharp distinct basal boundaries. Following the deposition of D2, the sediments of D1 would have been 'protected' from deformation. Conversely, the presence of large quantities of shale and other local bedrocks within D2 suggests that the two facies must have interacted in order for the shale to become part of this facies. Therefore the presence of a 'sticky spots' (Evans 2003) beneath the glacier could be inferred. This 'sticky spots' model was proposed in Piotrowski and Kraus (1997) suggesting that isolated spots of deformation could occur in between areas where ice-bed decoupling occurred, resulting in no deformation of the underlying sediments. Deformation occurred in areas where there was no significant reduction in ice to bed contact resulting in a 'mosaic of largely undeformed substratum with isolated spots of deformed bed' (Piotrowski and Kraus 1997 p 501). This implies at that areas up flow of Upgang the bedrock must have been ripped up and incorporated into the diamict, explaining the large quantities of shale by the process of advective excavation (Roberts and Hart 2005). Then, as the ice overran the Upgang area the basal conditions changed, possibly due to high contrast between the sediments or high strain in the upper medium (Boulton *et al.* 2001) or a change in strain rate (Alley 1991), resulting in décollement.

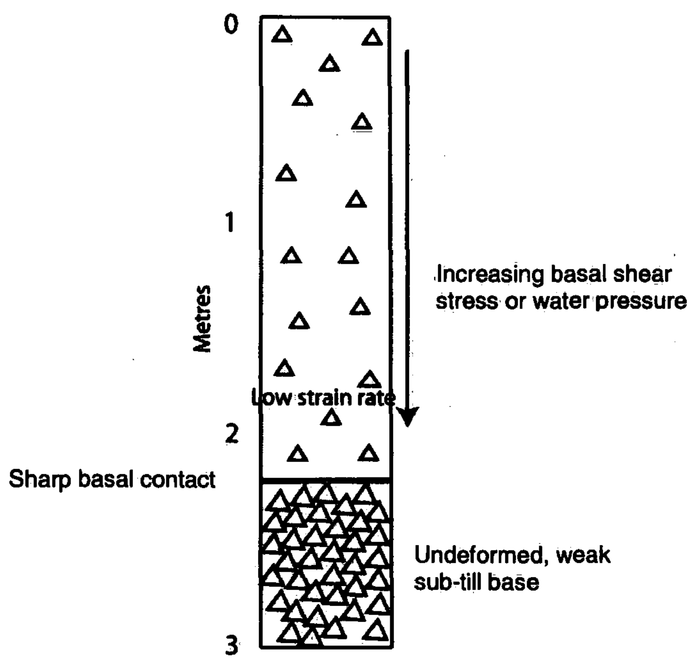


Figure 40: Annotated log from Section E showing the sharp basal contact between D1 and D2.

In one location D1 can be seen dipping below present day beach level, Section D (Figure 41). It is a gentle slope with the décollement plane continuing between the two units. However, as D1 slopes down towards the right a band of reddish diamict with very small clasts appears. This is potentially an injection feature formed as a result of differences in basal water pressure causing the red diamict to be forced up between D1 and D2, similar to a diapir formation (Menzies and Shilts 2002). The décollement plane between the two sediments would have acted as a structural weakness making this the easiest path for the red sediment to take. The red nature if the diamict is likely to have come from the Triassic sandstones from the north (Radge 1939). Streaks and rafts of red, along with grey and white material are also found along the Holderness coast suggesting incomplete subglacial mixing and shearing (Eyles *et al.* 1994).

Figure 41: Photograph showing the red injection feature in Section D. Spade for scale.

While, this evidence points towards the genesis of the till being a result of deformation, Evans *et al.* (2006) suggest that a number of processes lead to the formation of subglacial tills. Therefore, it should be classified as a subglacial traction till (Evans *et al.* 2006) which is dominated by deformation processes.

However, this facies of diamict is very thick compared to modern analogues for deformation tills found in Iceland, where the average thickness of till rarely exceeds one metre in height (Evans and Twigg 2003). It is possible that due to the topographical location of the site the till sequence has been thickened by ice marginal stacking. A till wedge is formed under a glacier with the thickest layer of deposits at the ice margin as a result of erosion and deposition patterns (Kjaer *et al.* 2003). Advance and retreat till is thought to withstand erosion, hence, increasing the quantity of sediment at a site (Kjaer *et al.* 2003). Due to Upgang's proximity to the North Yorkshire Moors, which were unglaciated during the last glacial, it is plausible that the ice margin was situated proximal to Upgang. Therefore, as ice pushed up against the uplands of the North Yorkshire Moors till would have thickened in the Upgang area. This could be as a result of till stacking, that has been shown to occur at the margins of modern analogues in Iceland (Evans and Hiemstra 2005). This process involves the incremental thickening of till by a combination of subglacial processes. As the till thickens the depth of deformation alters and results in the overprinting of earlier till horizons partially or totally eroding the horizons between layers (Evans and Hiemstra 2005). Hence, as the till thickens the lower deformation features are preserved and deformation style could be altered as a result of the changes in the substrate below this layer (Boulton 1996). It is due to Upgang's submarginal nature that this signal of both extensional and compressional deformation is seen. Material has been advected and transported through the zone of deformation into the marginal zone. In addition due to the sites location next to the North Sea the diamict will have been heavily saturated. It should also be noted that while modern analogues in Iceland show thin till sequences Truffer *et al.* (2001) present data showing a till sequence of 5 to 7 metres thick under the Black Rapids Glacier in Alaska, although this is a surging glacier. Additionally Dowdeswell *et al.* (2004) report thick layers of till under fast flowing ice streams of on average 4.6 meters thick in Antarctica. Therefore, it is also plausible that the thick vertical sequences of till were deposited subglacially without any stacking processes occurring. Alternatively, the thick diamict sequence could be attributed to the process of mass flow. These sequences can reach up to 10 m in thickness (Eyles 1987). Although, for a mass flow genesis hypothesis to be accepted in relation to D2 the other sedimentological evidence would need to match. In a mass flow unit clast fabrics would be expected to have high dip values and variable slope controlled fabric direction, whereas those taken from Upgang do not. In addition the eigenvalues, clast form data and many of the sedimentological and structural features preferentially support a deforming bed hypothesis. A mass flow unit would also be likely to display stratification representing various mass flow events. However, stratification is found at only a small section of the cliff, where it has been proposed the stratification was formed as a result of high shear strain values

Another feature of D2 is the colour change in the sequence separating D2A and D2B. D2A has a higher concentration of clasts of local bedrock. While D2B has fewer clasts of a smaller size than D2A with a marginally higher proportion of distally derived lithologies (Figure 26). This visual description, along with the suggestion that the clasts from D2A had a shorter transport distance from the C40/RA covariance plot, fits with the model proposed in Boulton (1996). This suggests that mean derivation distance of lithologies increases upwards within a sequence with sharp contrasts occurring across erosional interfaces. As the glacier overruns an area the local larger clasts are deposited first, then later ice with more distally derived smaller clasts will deposit further sediment. The boundary between the two units could represent two horizons of the till. Nevertheless, with the exception of the variation in lithology quantities and colour change, which can be explained by moisture content, the remainder of data fails to significantly separate the two units from one another. Similarly the geochemical data shows no real elemental differences between the two units, although, as only one sample has been analysed from each unit this cannot be seen as conclusive. This indicates that D2A and D2B are of the same unit but reflect the increasing distally derived sediment up throughout a sequence.

The provenance of this facies can be determined from the orientation of the clast fabrics and from clast lithology data. The rose diagrams (Figure 20 and 21) suggest ice flow was from north-north-west. While a large number of these diagrams using the a-axis orientation suggest ice flow from the north east these could be lying transverse to flow as a result of deformation (Hart 1994). Clasts do not always align parallel to flow with some lying transverse to flow and particles of different sizes show different rotations as each particle reacts in response to the individual stresses placed upon it (Carr and Rose 2003). Data was taken on clasts in a similar size range (8 - 32 mm) which will minimise the differences in rotation dependent on size. The a-b plane data indicates that ice flow originated a northerly direction as these rotate in line with glacier flow more readily than a-axes (Evans *et al.* 2007), supporting the a-axis fabrics that also indicate a north-north-west origin for ice flow.

A northern origin of the ice is supported by the clast lithological counts. As suggested in Section 4.4 many of the erratics found within the till can be traced to the Cheviot Hills, Northumberland and County Durham, along with the presence of flint from the North Sea. Combining this data with the ice flow directions it can be proposed that the till was deposited by the North Sea ice lobe that existed along the coast of Yorkshire during the last glacial.

5.3 Sands and Gravels

The sands and gravels layer has been split into four units, Section 4.2, to aid description and interpretation. Therefore, each unit will be discussed separately before being collated to consider the facies as a whole. It is proposed that this entire sands and gravels section was deposited in a lacustrine and not a marine environment. The sediments that make up the section lack any shells or forams that would suggest that they were deposited in a marine environment. In addition, as a marine environment would be tidal the presence of herringbone cross stratification and flaser bedding would be expected especially in the finer sediments. However, this is also absent from the sequence.

5.3.1 SG1

SG1 is a facies consisting of laminated clays, silts and fine grained sands which can be seen at the base of each of the sections. The laminated fine nature of these sediments, along with the lack of shells or forams, indicates that they were deposited in a lacustrine environment. As the sediments are fine, water velocity would have been low or ponded with sediment deposited as suspension rain out (Brazier *et al.* 1998). It is possible that the finer grained sediments within this facies were deposited during winter when flow into the lake decreased as a result of reduced melting of the ice or the lake froze over allowing them to settle out (Teller 2003), although there does not appear to be a coupling of finer sediments and sand that could be classified as varves representing seasonal changes in sedimentation. The thickness and nature of these laminated sediments could be a result of sediment supply, depth of water and proximity to water source, as opposed to seasonal variation (Teller 2003).

The coarsening upwards of this facies could indicate the increasing proximity of an ice mass to the area (Smith 1985). The rare presence of Type B climbing ripples in the sand towards the top of the unit supports this indicating a reduction in suspended sediment load suggesting that the lake was beginning to shallow (Teller 2003). These ripples along with the coarsening upwards nature of the section could also indicate the initiation of, or an increase in bottom currents and underflow activity, supporting the hypothesis of a shallowing environment.

5.3.2 SG2

SG1 continues to coarsen upwards grading into a facies of laminated sands containing more frequent ripples and some planar bedding, SG2. The coarser nature of the sediments could imply a switch from a deeper water environment to sedimentation in shallower water (Benn and Evans 1998) as an ice mass approached the area. Alternatively, this could represent increased bottom current activity. It is the increasing coarse nature of these sediments which suggest that the ice is getting closer to the lake, although slope instabilities and river channels could also play a role. Alternatively, they could represent increased current activity. The Type B ripples found within these sand units record deposition from suspension of the whole bedform (Allen 1973), indicating turbid underflows and high aggradation rates (Ashley *et al.* 1982). These ripples also represent an increasing bed load coupled with decreasing suspended sediment indicative of an increasingly shallow environment (Teller 2003).

Paleocurrent directions from this facies in Section Bii (Figure 9a) indicate that water flowed from approximately the south-west, while in Section J (Figure 15) they infer that flow direction was from the north east. This indicates that in Section J the water is flowing away from the advancing ice margin, while in Section Bii it is flowing towards it. Therefore, water is originating from either an advancing ice mass to the north in the case of the flow from the north east, or that it is run off from the North Yorkshire Moors to the south west. The conflicting directions of flow could be due to meandering streams that deposited the ripples.

The occurrence of units of matrix supported granules within Section Bii and Section J indicate periods of more rapid flow or a further lowering of water levels. Conversely, the clay unit towards the top of Section Cii suggests a period of water ponding. Also of interest within Section J is a near vertical elongate gravel pod within the rippled sands (Figure 15). This pod could be the relic of pipe flow within the sand. This type of feature is common where the bed is unconsolidated (Benn and Evans 1996). Pipe flow is an element of water drainage, typically subglacial. This pipe may have formed either as a drainage feature when the overlying gravels were deposited or when ice overrode the site. Although if deposited subglacially the pipe appears to be at a considerable depth to the ice itself. Alternatively this sub-vertical gravel pod could be a dewatering feature representing the disruption and displacement of sediment as a result of escaping pore fluid (Benn and Evans 1998). Furthermore, within Section J, above the rippled sands, is a unit of cross bedded sand foresets. These could either be Type A ripples resulting from high stoss side erosion or could have been deposited as a result of dune or bank migration (Benn and Evans 1998). As well as representing the infilling of a lake and the reduction in water velocity these features could also indicate that the area was becoming increasingly ice proximal. This facies does not appear to exist in Section D, with the laminated sands and fines of SG1 appearing to grade into the granules and gravels of SG3 without hiatus.

5.3.3 SG3

The coarsening upward of the sands and gravels unit continues leading into SG3. This facies consists of granules and gravels with some interbedded laminated sands. The occurrence of the larger sized particles indicates deposition by faster flowing shallow water in the upper flow regime (Maizels 1993). The cross trough bedded shape of the deposits suggests that this was in channels in an increasingly ice proximal setting in front of advancing ice (Thomas 1987). Laminated sand pods found within gravel units in Section Bii could be rip up features, with water dissecting the laminated sands below and incorporating them within the gravels (Cornwell 1998). The nature of these gravel and granule channels also suggests that they were infilled rapidly. As they are situated in unconsolidated material, a slow infilling mechanism would have caused the slopes to be degraded by slumping (Shaw 1987). Till balls found within this layer also indicate that the deposition of till has already started to occur up ice flow as the gravel channels were being deposited at Upgang (Shaw 1987). Alternatively, these till balls could be derived from the erosion of D2 below. However, their colour is very similar to the colour of D3 and not that of D2.

The direction of paleocurrents taken on the foreset bedding and laminated sands of SG3 in Section Bii indicate predominant flow directions from the south with various units showing flow directions from the north east (Figure 9a). Again it is possible that the water originated as meltwater from the advancing ice, which is believed to have originated from the north to northwest from clast fabrics taken within D3, Section 5.4, but that the channels are meandering out in front of it. This causes the appearance of water flowing in the 'wrong' direction especially for the directions taken from the gravel bedding. In meandering and braided rivers, 'bedforms migrate at a variety of angles to the main trend of the river' (Benn and Evans 1998 p 402). The flow directions within the sand units could have been altered as these units may be a result of rip up within the channels while the remaining directions indicating flow towards the north have been taken from foreset bedding. As discussed above these could represent dune migration and hence, would not necessarily record the direction of main river flow. However, an alternative proposition, especially for the currents from the south, is that the water originated from the North Yorkshire Moors, as three modern day becks enter the sea from the North York Moors at Upgang and Sandsend. Although as the sequence of SG3 indicates an increasingly ice proximal location, it would seem more likely for the water to have originated as proglacial meltwater flows.

5.3.4 SG4

SG4 represents a thin layer of laminated and rippled sands at the very top of the facies, overlying the gravels of SG3. This facies can only be seen in Section B and Section J, as the upper boundary of the sands and gravels is only seen clearly in these locations. At Section C and D the boundary is obscured either by slumping or it was not possible to gain access to consider the site in detail. As this is a return to finer material after the deposition of the upwardly coarsening units, it could be interpreted as the deposit of a sheet flow feature that can form at the ice-sediment interface or within till. This implies that the overriding ice mass was warm based and that meltwater would have exceeded the amount that could have been evacuated by advection through the till or Darcian through flow (Murray 1997). As a result discrete channels could have formed at the interface between SG3 and D3, as a result of a structural weakness between the two, leading to the deposition of SG4. As the ice overriding the Yorkshire coast was part of an ice lobe rather than a valley glacier these channel should be low and wide (Walder and Fowler 1994). The shape of SG4, which can be seen at the top of Figure 9, shows that the channel would have been both wide and shallow, strengthening this sheet flow hypothesis. The partial laminations and ripples within SG4 also support this model as the ripples indicate the presence of flowing, potentially turbulent, water, as substantiated by the variable paleocurrent directions. However, this feature could also represent a wide shallow proglacial channel that formed in front of the approaching ice mass. Due to the nature of the sediments within

SG4 this channel appears to have been low energy, with the ripples representing underflow activity.

5.3.5 Whole Facies

Overall this sequence of coarsening up sands and gravels represents the infilling of a lake as an ice mass advances (Figure 42). The pattern of deposition is similar to that found in other areas that have been described as ice proximal lakes that grade upwards into deltas and sandur plains, for example Lairig Ghru and Gleann Einich in Scotland (Brazier *et al.* 1998).

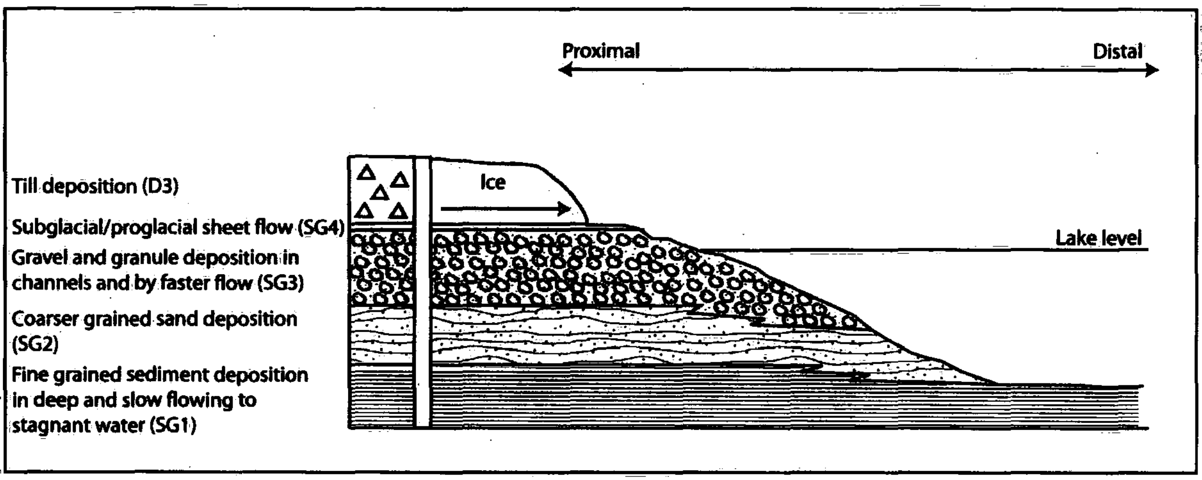


Figure 42: Cartoon simplifying the deposition of the sands and gravels at Upgang.

5.3.6 Section A

Also logged within this facies is the large fines unit found within Section A. Section A appears to belong to the upper part of SG2 with the lower part of the section dominated by laminated and trough bedded sands with some granules. However, the middle third of the section is a unit of fines including laminated sand deposits before an abrupt start to a unit of sands and granules. This fines unit has not been seen in any of the other logged sections. It is likely that this unit of massive fines was deposited as a result of sediment settling out in ponded water. The isolated circular, oval and concave elongated sand pods within the fines unit (Figure 43) could be a product of gravity induced soft sediment deformation. As a result of saturation and high water pressures sand pods can be deformed in to a sequence of fines (Harris *et al.* 2000). High water pressure would have resulted in the liquefaction and failure of the sand. This sand-water mixture is then injected into the fines unit following a path of least resistance (Visser *et al.* 1984). These involutions form circular and polygonal structures (Murton *et al.* 1995) similar to those seen in Section A. The isolated nature of these features infers that the overlying sand was deposited rapidly resulting in the loss of strength within the fines unit (Collinson and Thompson 1992). This type of gravitational soft sediment deformation often has a sharply defined basal boundary (Visser *et al.* 1984) as can be seen in Figure 43.

Figure 43: Photograph of part of Section A showing soft sediment deformation. Several of the sand inclusions have been highlighted. Trowel for scale.

5.4 Diamict 3 - genesis and provenance

Due to the similarities in the quantitative data between D2 and D3 it is sensible to suggest similar depositional histories, with the facies also being described as a subglacial traction till influenced by deformation. The subtle differences between D2 and D3 are a result of the differences in underlying sediments. As the two diamicts are the products of two separate ice advances, the substrate underlying each ice advance is different; underlying D2 is the solid glaciotectonic unit of D1, while below D3 is the sands and gravels facies. Therefore, differences will have occurred in the subglacial effective pressure system (Roberts and Hart 2005).

The clast form data suggests a subglacial origin to the sediment. D3 plots lower on the covariance plot (Figure 34) than D2, further towards the corner of the graph that was assigned as having a subglacial transport history (Benn and Ballantyne 1994). The clast fabric data also indicates that the sediment has been subglacially deformed, with points plotting in the deformation section of the modality-isotropy diagram (Figure 35). Although this, along with the step contoured stereonets, implies that there could be increased lodgement higher up the sequence, conforming to the hypothesis of a subglacial traction till. The eigenvalue plots also confirm this hypothesis of deforming processes resulting in the deposition of the diamict facies. The position of the samples from D3 on the ternary and S1/S3 plots vary only subtly from D2 (Figures 36 and 37) plotting within the envelopes assigned to deformation tills (Benn, 1995; Hicock *et al.*, 1996). However, the validity of these envelopes has been questioned along with the reliance on clast fabric data to indicate till depositional histories (Bennett *et al.* 1999; Carr and Rose, 2003).

As with the clast lithology data from D2, D3 displays a high percentage of locally derived clasts. However, as Figure 26 shows this facies has a higher number of distally derived lithologies than D2. The relatively low percentages of shale are a result of D2 'capping' the shale bedrock locally, preventing the rip up of clasts pertaining from the original bedrock. Instead many clasts within this facies will have been taken from D2 and SG.

Another feature that indicates a subglacial origin to the sediment is the sand pod found in Section B (Figure 44). This pod appears to have been pinched at either end which can suggest boundinage by high shear stresses (Hart and Roberts 1994). This pod maintains much of its primary structures in the form of laminations, although there appears to be some folding of these laminations towards the middle of the feature. Benn and Evans (1996) described similar features at Loch Lomond, Scotland inferring that the pods acted as stiff regions within the unit being 'rafted along with more rapidly deforming finer grained units' (Benn and Evans 1996 p 41), indicating that the feature was a result of rip up. The clast fabric taken to the right of the feature (sample 3.1, Figure 21), shows higher than average fabric strength for the site. Equally, this feature could be an infilled englacial channel that has later been deformed and pinched by the stresses placed upon it. This would favour the model proposed by Piotrowski *et al.* (2001) that sand pods cannot survive in a deforming bed unit, although when considered alongside the other evidence from D3 a deforming bed origin is more likely.

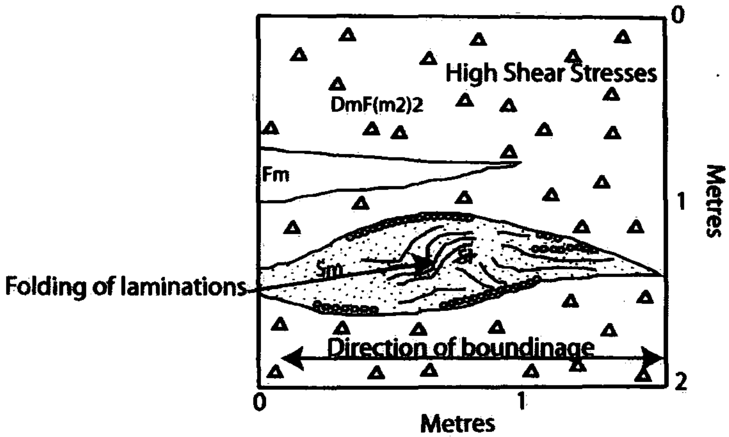


Figure 44: Annotated sketch of sand pod in LFA D3 in Section B.

Also within Section Bi is an apparent rip up feature consisting of a laminated sand unit, probably taken from the underlying sand and gravel section (Figure 8b). This feature appears to have been ripped up and deposited within D3 without the internal structures being considerably altered, although a fault does appear to exist to the left of the feature. It is difficult to find another explanation for the existence of this feature although only the upper boundary has been exposed, meaning that other indicators of its depositional history have been hidden. Other sand inclusions are also found towards the top of Section Ci (Figure 10). Due to the shape of these features it is possible that sand has been deposited within the till and then folded and deformed (Hart 1994). The clay fold towards the top of this section indicates a similar origin.

The shear stresses that appear to have been present within D3 could also account for the mixing of clay within the diamict at the top of Section Di (Figure 11). Low strain values within D3, which would also account for the preservation of the sand and clay features, could cause the folding and partial mixing of the sediment without homogenising it, as mixing is proportional to strain (Hooyer and Iverson 2000). However, this mixing can also be seen as a result of a mass flow deposition history. Shaw (1987) described a heterogeneous diamict that is mixed with sorted sediments. This was considered to have occurred as a result of the intermixing of the two sediments by a combination of processes during mass flow. Nevertheless if this mixing was a result of mass flow processes then the clast fabrics should exhibit high dip values with variable flow directions, which they don't. In addition, the presence of stratifications representing flow units would be expected yet there is no stratification found within D3.

The basal contact of D3 with the sands and gravels facies is only visible in one small section of the field site at Section Bii. Here the contact appears to be fairly distinct, although there is some slight mixing of sand and diamict along with the deposition of diamict rafts within the sands (Figure 45). This boundary can be indicative of a deforming bed unit as many authors believe that the lower sediments can become deformed as the ice and till override them (Hooyer and Iverson 2000). This would account for the mixing of the sand within the diamict. As with the basal boundary of D2, this boundary would also require areas of larger scale erosion to account for the sand pods and rip up features within the facies. Another explanation for the sand at the base of the diamict and the rafts of diamict within the gravels is the presence of water at the ice-sediment contact as the ice initially overran the area. Beneath a warm-based ice mass it is believed water could be drained by sheet flow as the amount of melt water will exceed the amount that can be advected through the till (Murray 1997). Thus, a system of discrete channels could exist between the ice-sediment interface or within the till. This would account for the deposition of the very upper sand facies (SG4) above the underlying coarsening upwards sequence of SG1 to SG3 before the deposition of D3. Although this facies could also be the result of a wide shallow proglacial channel. It is therefore possible that this water caused the detachment of the diamict rafts which were then pushed down into the sand and gravels. Another explanation leading to the deposition of diamict below its stratigraphic layer could be the 'squeezing down' of sediment into depressions within the lower substratum due to the pressure of overriding ice (Ruszczynska-Szenajch 1987). This could account for the presence of the rafts of D3 within SG4 regardless of whether SG4 was proglacially or subglacially deposited.

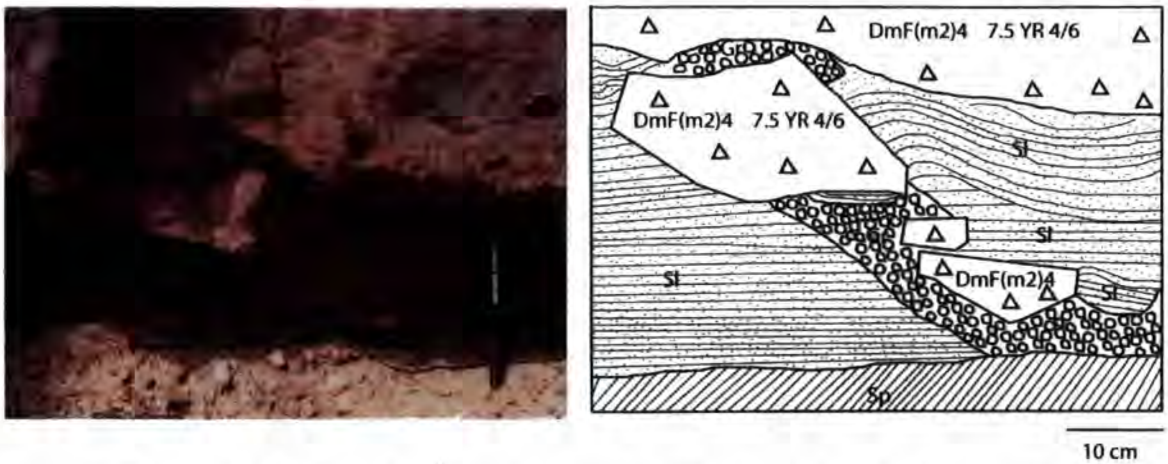


Figure 45: Photograph and sediment log of diamict rafts within SG unit at the boundary between SG and D3.

As discussed above, some modern analogues only show thick till sequences where the till has been stacked (Evans and Twigg 2002, Evans and Hiemstra 2005), while others are 4.6-7 meters thick (Truffer *et al.* 2001, Dowdeswell *et al.* 2004). The vertical extent of D3 exceeds six metres in places, and thus, may imply the need for a mechanism to exist to emplace such a thick sequence. Therefore, the process of till stacking is again proposed to create this facies. The stacking of tills does not necessarily imply that banding will be found within the unit, with many stacked tills being homogenous in nature. In order for this process to occur, Upgang would have to be ice marginal as discussed in section 5.2. The geomorphological map from the NEXT Map data (Figure 31) shows that the North Yorkshire Moors were not glaciated, and due to the proximity of Upgang to this area it is plausible that it was ice proximal.

The provenance of the facies is similar to that of D2. The clast fabric data of both a-axis and a-b planes indicate that ice flow originated broadly from the north west. Lithologies from the diamict also indicate a northern origin with distinct clasts of pink rhyolite from the Cheviot Hills. Along with coal, dolerite from the Whin Sill, greywacke and Carboniferous Limestone indicating that the ice travelled from the north of England and Southern Scotland. The red colour of the diamict is thought to originate from the Triassic Marls of the Tees Estuary (Radge 1939). This data suggests that the second diamict at Upgang was also deposited by ice from the north or the North Sea as the Triassic Marls extend into the sea (Figure 4). The red colour of this till could have been emphasised by a weathering profile caused by oxidation, as is seen on the Northumberland coast (Douglas 1991).

6. Discussion: Correlation with Yorkshire, climate and LGM readvances.

6.1 Is there a readvance signal at Upgang?

From the discussion of the glacial sediments at Upgang in Chapter 5 an advance-retreat-readvance succession can be proposed. D1 and D2 were deposited as part of the initial advance phase. The lower diamict D1 represents either a glaciotectonite formed from the Jurassic bedrock underlying Upgang or an overrun proglacial thrust feature, while D2 is a subglacial traction till dominated by deformation processes. The overlying sands and gravels LFA then represents the retreat of the ice resulting in the formation of an ice proximal lake. Ice then readvanced into the area leading to the infilling of the lake. This is supported by the increasingly ice proximal nature of the sands and gravels unit along with other stratigraphic indicators. This includes the presence of till balls, which have been correlated with D3, within SG3 indicating that the deposition of this till had begun up flow at the time the gravels were deposited. The deposition at the top of the sequence of another subglacial traction till, D3, implies that ice then readvanced over the whole area. The entrainment of the lacustrine sediments below into D3 supports this readvance hypothesis.

Therefore, the sedimentary sequence at Upgang clearly shows ice advance, D1 and D2, ice retreat leading to proglacial, lacustrine and subaerial sandur sedimentation, followed by an ice sheet readvance depositing D3.

6.2 How does Upgang fit into Yorkshire's glacial history?

However, this advance-retreat-readvance interpretation is not consistent will published work from the Yorkshire coast. The limited research previously carried out at Upgang has suggested a number of glacial histories for the area which contradict the new proposal outlined above. Harrison (1895) suggested that the triple division of the section could be attributed to a single ice advance. The lower till was deposited and compacted by the initial advance. The sands and gravels then represented englacial channel fills while the upper till was believed to be the result of the final melting of the ice. The infilled lake (sands and gravels LFA), along with its lateral extent contradicts this proposal. The upper till is also better described as a subglacial till rather than a melt out till. Madgett and Catt (1981) claimed an intermediate till bed between the 'Skipsea' and 'Withernsea' till was found at Upgang. This was said to represent a zone of mixing in a multilayer ice sheet. However, this layer has not been found during this investigation weakening the argument for a multilayer ice sheet. Catt (2007) provided the most extensive published history of the sediments at Upgang. The lowest grey till, D1, was deemed to be a pre-Late Devensian deposit due to the presence of a weathering profile in the upper section of this LFA along with the sharp erosional boundary separating it from D2. Although no other evidence for its age was presented, on the basis of its apparent weathered edge it was correlated with the Basement Till of Dimlington. This weathered till was not witnessed in this study. Despite this proposal this investigation interprets this LFA as a Devensian glaciotectonite or overrun proglacial feature on the basis of the evidence in Section 5.1. The remaining sediments were again correlated with the Skipsea and Withernsea tills of Holderness, although the means of deposition was not discussed.

The new interpretation of the glacial history of Upgang presented in this study is also inconsistent with the published work from Holderness. Along this coast two tills have been identified, the lower Skipsea Till and upper Withernsea Till separated by a distinct boundary (Catt 2007). No laterally extensive sorted sediment unit is seen. These tills have been correlated both north and south of Holderness. However, the provenance and genesis of these tills has lead to some controversy. Therefore, the glacial sequence of Upgang needs to be considered within this context in order to establish how it correlates with the sequences to the south.

6.2.1 Provenance of tills and ice flow patterns in the area.

The provenance of the Skipsea and Withernsea tills along the Yorkshire coast has proved controversial. One till was assigned to ice arriving from the north of England and southern Scotland, while the other has arrived through the Stainmore Gap from the Lake District. However, authors cannot agree on which till originated from which location. Bisat (1939), Radge (1939) and Foster (1987) suggested that the Skipsea till was from the Lake District with the Withernsea till originating in the north, while Catt and Penny (1966), Madgett and Catt (1978) and Catt (2007) proposed the opposite. These studies were all based on clast lithological studies with local ice flow directions being taken from clast fabric orientations.

However, the clast lithological data from Upgang indicates that the two tills present at this site contain similar erratics and therefore are likely to have originated from the same source. Due to the proportionally high percentages of greywacke (Southern Scotland), dolerite (Whin Sill), Carboniferous Limestone (Northumberland), coal (County Durham) and igneous rocks from the Cheviots, ice flow is from the north of England. Erratics from the North Sea (flint and chalk) are also found in both tills to support a proposal that both tills are deposits of a North Sea Ice Lobe. It is possible that erratics sourced from the Lake District described by previous authors were deposited in the area prior to the Devensian and have been incorporated into these tills. Clast fabric orientations showing ice flow onshore from the North Sea at Upgang support this hypothesis along with clast lithological data from North Norfolk (Pawley *et al.* 2006). The ice arriving at Norfolk would have had to overrun Yorkshire due to its ice flow trajectory. This shows ice flow only from north east England and southern Scotland, however, only one late Devensian till is present in Norfolk.

The geochemical data taken from Upgang can also be used to support this hypothesis. Samples taken from D2 and D3 show no significant difference in elemental composition, with relatively low chi-squared values and the cluster dendrogram showing clustering of samples between units as well as within (Figure 27). While these results do not show the same composition within samples, the differences are not great enough to indicate separate sources for the two tills.

In order to support this hypothesis of ice flow direction solely from the Cheviots, indicators for the wider area also need to be considered. The NEXT map data (Figure 32) along with the BRITICE data (Figure 31) shows the ice flow patterns in North East England from lineations and moraine ridges. Ice crossed from west to east in two locations within this area, the Tyne Gap to the north and the Stainmore Gap. The lineation patterns from these areas indicate that ice from the Tyne Gap continued towards the North Sea and was deflected south. While ice from the Stainmore Gap appears to have been deflected southwards over a shorter spatial area. This suggests that ice from the north may have been present over the Tees Estuary and coast in order to cause this deflection. Ice from southern Scotland and northern England is thought to have arrived at the east coast along the north east coast and through the Tyne Gap, with Lake District ice forced through the Stainmore Gap. Therefore as the ice from this area appears to have been blocked from entering the North Sea by other pre-existing ice lobes it is unlikely that Lake District ice from the Stainmore Gap was present along the Yorkshire coast. Instead it would have been forced into the Vale of York. In addition, if the Tees Estuary area was unglaciated allowing the Lake District ice to flow along the Yorkshire coast towards Dimlington a mechanism would be required to deflect the ice southwards rather than simply spilling into the North Sea. Initially, it was suggested that Scandinavian ice must be present in this area causing the ice to deflect, however, the presence of Scandinavian ice during this time has recently been questioned (Serjup *et al.* 2000). The proposed presence of a glacial lake in the Tees Estuary (Agar 1954) also questions the validity of Lake District ice reaching the east coast. For this lake to exist it must have been dammed by ice from the North Sea Lobe to the east and ice entering the Vale of York to the west. Therefore, Lake District ice would not have been able to flow through the Tees valley to the coast. Nevertheless, Catt (1991) suggested that this Lake District ice was forced on top of the North Sea Ice Lobe in the Tees Estuary which led to the model of a multi-layer ice sheet.

However, erratics found at Upgang include those originating from the Cheviot Hills. This ice must have flowed from this area in order for these rocks to be incorporated within the sediment. This would imply that ice was predominantly from the north flowing down the coast. However, the possibility of sediments being reworked from underlying diamicts must be considered due to the low percentage of erratics. Ice flow patterns in this area are highly complicated, increasing the potential for reworked sediments. Interpretations are hindered by the lack of knowledge regarding the extent of the Scandinavian ice sheet and the glaciation of the North Sea.

Therefore, using the above evidence, it is proposed that both tills at Upgang have the same northern source and were both deposited by a North Sea Ice Lobe. It is also suggested that the Skipsea and Withernsea tills are also from this same northern source; however, this is beyond the scope of this project.

6.2.2 Mode of deposition

Unlike Upgang, at sites in southern Yorkshire where both Late Devensian tills outcrop, the Withernsea Till directly overlies the Skipsea without any laterally extensive sorted sediments. Many different mechanisms have been discussed regarding the deposition of these deposits including surging (Eyles *et al.* 1994), a dual layer ice lobe in a single advance (Madgett and Catt 1978) and vertical till accretion (Evans *et al.* 1995). The two tills appear to have been differentiated on colour difference and differences in clast lithology. It has been suggested above that it is possible both tills originated from the same location and, hence, show similar lithologies leaving only a colour difference between the two tills. Therefore, there may only be one till with differences in colour stemming from different underlying bedrock along with differential mixing and folding of the till. Although, work would need to be carried out in order to verify this hypothesis. This suggests that the Skipsea and Withernsea tills represent a stacked sequence and do not represent two separate ice advances or a multilayer ice sheet. Evans *et al.* (1995) suggested that vertical till accretion of a deforming layer was responsible for the till sequence at Filey Brigg a sequence, which Catt (1991b) had previously interpreted as displaying both Skipsea and Withernsea Till. This further supports the suggestion that the Skipsea and Withernsea Tills are a stacked sequence and indicates that D3 at Upgang may not correlate with the Withernsea Till.

It is also possible that the glacial sequence of Holderness was deposited as part of a surging ice lobe as proposed by Eyles *et al.* (1994). While the tills along Holderness may have been deposited as part of a surging lobe it is highly unlikely that those at Upgang can be correlated with these surges. While D2 could correspond to the surge depositing the Skipsea Till, Upgang is considerably further north than the area in which the Withernsea Till outcrops. Hence, the ice lobe must have retreated further north than Upgang before the surge that deposited the Withernsea Till to allow for the deposition of D3. However, this would indicate that the extent of the Withernsea till would need to be more expansive. Although a paleao-surging landsystem is difficult to identify, the presence of large thrust and push moraines along with patchy hummocky moraines would be expected if glacial surging had occurred, following the model proposed in Evans and Rea (2003). The NEXT map data (Figure 32) does not appear to show such features.

6.2.3 Extent of the readvance

If this hypothesis of a single till unit for the majority of East Yorkshire is accepted then it can be suggested that the lower tills at Upgang correlate with this initial advance of ice along the coast. Ice then retreated to a position further north than Upgang before readvancing. However, the extent of this readvance would need to be examined as the limit would exist somewhere between Upgang and the Holderness coast. Proglacial lake sequences have been shown to exist at Barmston and Skipsea along this coast (Evans *et al.* 1995). Although there is no evidence of glacial deposits overlying these, suggesting that the limit lies north of Barmston. Nevertheless, the ice lobe readvance may have travelled south of these sites remaining offshore and, hence, not overrunning them.

Alternatively, the sequence found at Upgang may be a response to local conditions, with ice retreating and readvancing or surging locally and not along the whole coast. Tills deposited by a surging glacier produce similar glaciotectonic features and the thickening of the sediments to the flow of non surging glaciers over a deformable bed (Evans and Rea 2003). Therefore the upper till found at Upgang could have been deposited by a non surging glacier overriding a deformable bed or a surging glacier. However, paleao-surging can be difficult to identify as no landform or sediment sequence unequivocally identifies this style of glaciation (Evans and Rea 2003).

<u>6.2.4 Summary</u>

Therefore, as the sediments at Upgang show an advance-retreat-readvance succession the glacial history of north Yorkshire needs to be reassessed. Clast lithological analysis suggests that both tills were deposited by a single northern ice lobe. However, current interpretation of the glacial sequences at other sites in Yorkshire does not support this. While the apparent proglacial lake deposits capping the sequence at Barmston and Skipsea (Evans *et al.* 1995) suggest a readvance limit to the north of this area, the presence of the Withernsea till to the south could be contradictory. In order for this hypothesis of a single North Sea ice lobe depositing both tills, either by surging, till accretion or retreat and readvance to be accepted, the coastal sections to the south of Yorkshire need to be re-examined.

6.3 Can the glacial history of Upgang be linked to climatic forcing? Vegetation and climate records for North Yorkshire.

The sediments of Upgang have been interpreted as an advance-retreat-readvance signal. The mechanism of this sequence could be a result of ice surging, influenced mainly by internal dynamics or a result of changes in climate, with amelioration followed by cooling. In order to test if climate was a major driving factor in the readvance of ice at Upgang, published proxy records have been examined. If significant changes in climate in the area can be found then a link between the ice advance and climate can be considered.

Two key sites in the reconstruction of climate in northern Yorkshire are Kildale and the Tees Estuary (Figure 46). The proxy records from these sites are shown in Figure 47. The record from Kildale (Innes 2002) contains one radiocarbon date which has been calibrated using Calib Rev 5.0.1. The vegetation graph has then been 'wiggle matched' with the Greenland ice cores in order to provide a longer chronostratigraphic framework. The Tees Estuary record was 'wiggle matched' to the GISP ice core record in Plater *et al.* (2000). The British climate records at the end of the last glacial show a strong correlation with the climate records from the Greenland ice cores (Mayle *et al.* 1999). Thus, the Greenland ice cores can themselves be used as a proxy for British climate as well as providing a continuous record for other records to be compared to.

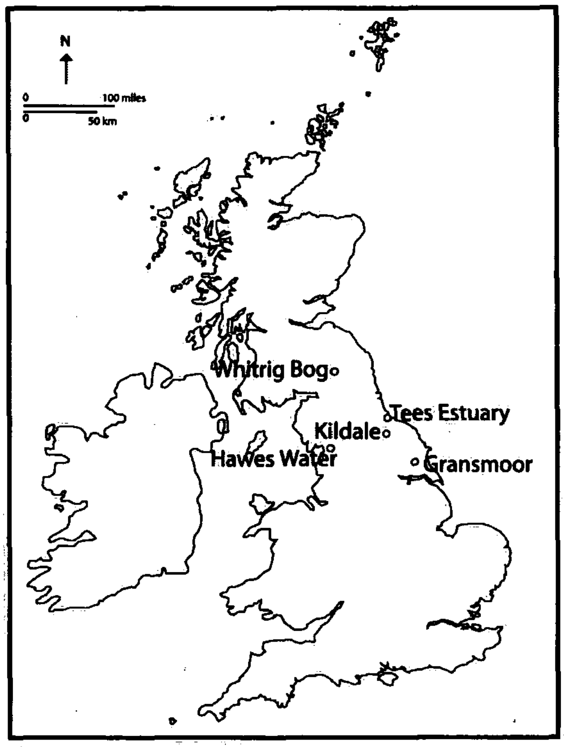


Figure 46a: Map showing the location of the proxy records listed in Figure 47.

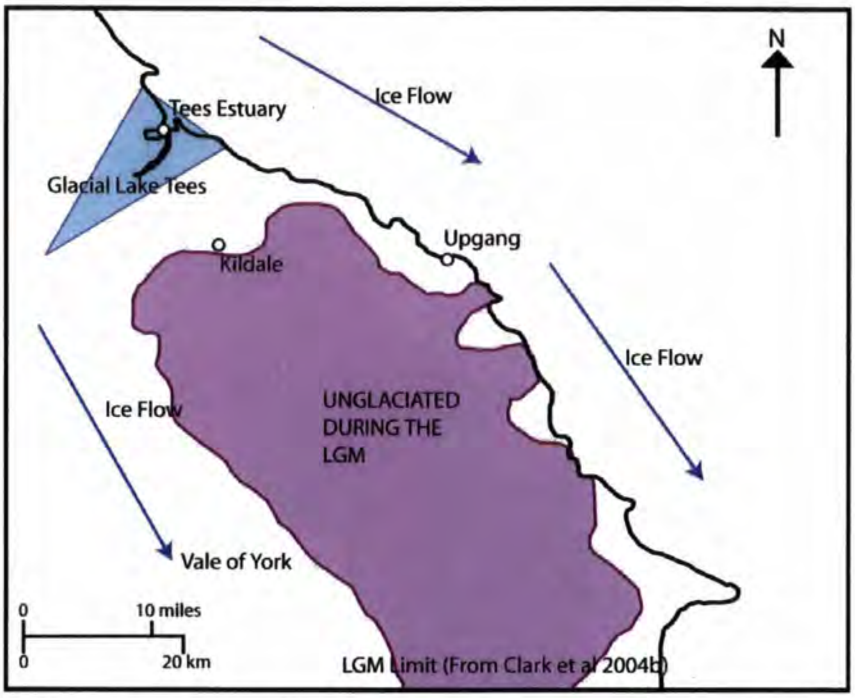


Figure 46b: Map showing the location of Kildale and the Tees Estuary with respect to Upgang, the LGM limit and Glacial Lake Tees. Glacial Lake Tees formed during the deglacial phase of the LGM between 18 000 and 16 000 cal. years BP. It was formed by meltwater from the surrounding ice masses (Agar 1954).

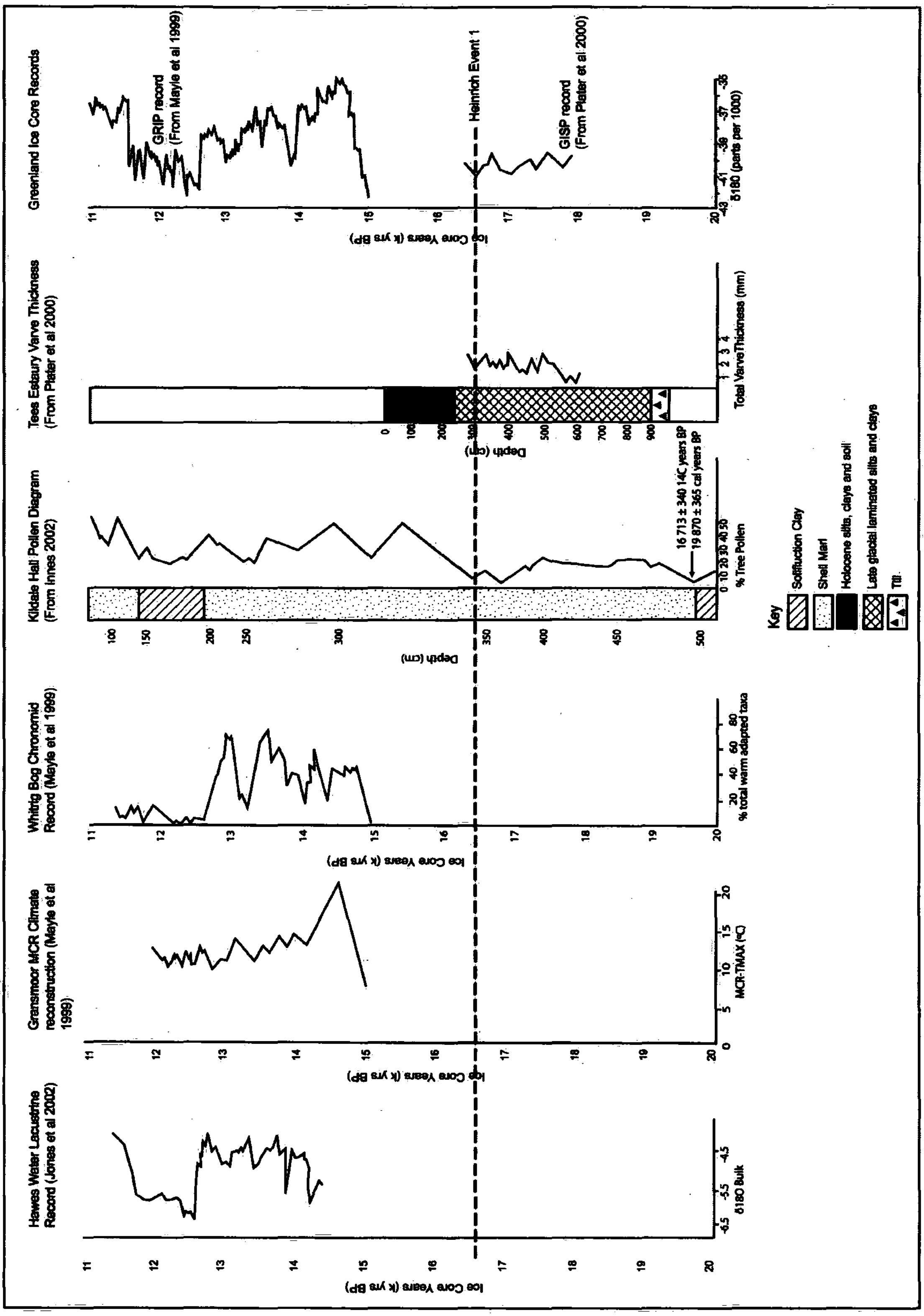


Figure 47: Proxy record graphs from Hawes water (Jones *et al.* 2002), Gransmoor (Mayle *et al.* 2000), Whitrig Bog (Mayle *et al.* 2000), Kildale (Innes 2002), the Tees Estuary (Plater *et al.* 2000) and the GRIP (Mayle *et al.* 2000) and GISP (Plater *et al.* 2000) ice core records.

6.3.1 Kildale

The vegetation record from Kildale in North Yorkshire provides a key date for deglaciation of the east coast, 19 144 – 20 537 cal. years BP (Jones 1977). Due to the position of this date within the core the vegetation record above it can be used as a proxy for temperature at the end of the last glaciation. The x axis on the graph (Figure 48) shows the percentage of tree pollen within the core; the greater the percentage the warmer the temperature. A deterioration in climate is shown at a depth of around 348 cm, zone KA-3 as assigned by Innes (2002) (Figure 48). By comparing this graph with the Greenland ice core records, a chronostratigraphy can be created suggesting that this cooling occurred between 17 400 and 16 500 cal. years BP. After this cooling event the climate beings to ameliorate with increases in tree pollen, although it is still unstable. A second significant cooling period is seen in KA-5, which correlates with the Loch Lomond Stadial (LLS). These changes in the percentage of tree pollen are seen more clearly in Figure 48, as Figure 47 has an exaggerated y-axis in order to match it to the ice core records. The lithostratigraphy beside the graph shows that during this time Kildale was not glaciated. However, as it is positioned just within the LGM ice limit it is in a good position to reflect the climate influencing any local ice mass. Although as Yorkshire was in a 'snow shadow' (Harrison 1895) it is unlikely that changes in the local climate would have greatly affected the ice mass. Instead it is climatic changes in the potential source areas of Scotland and the Lake District would have influenced that ice mass to a greater extent. Nevertheless, an advance of ice into the area would cause a drop in local temperature. Therefore, while the mechanism of advance cannot be confirmed by the record at Kildale, an approximate timing for the advance can be inferred.

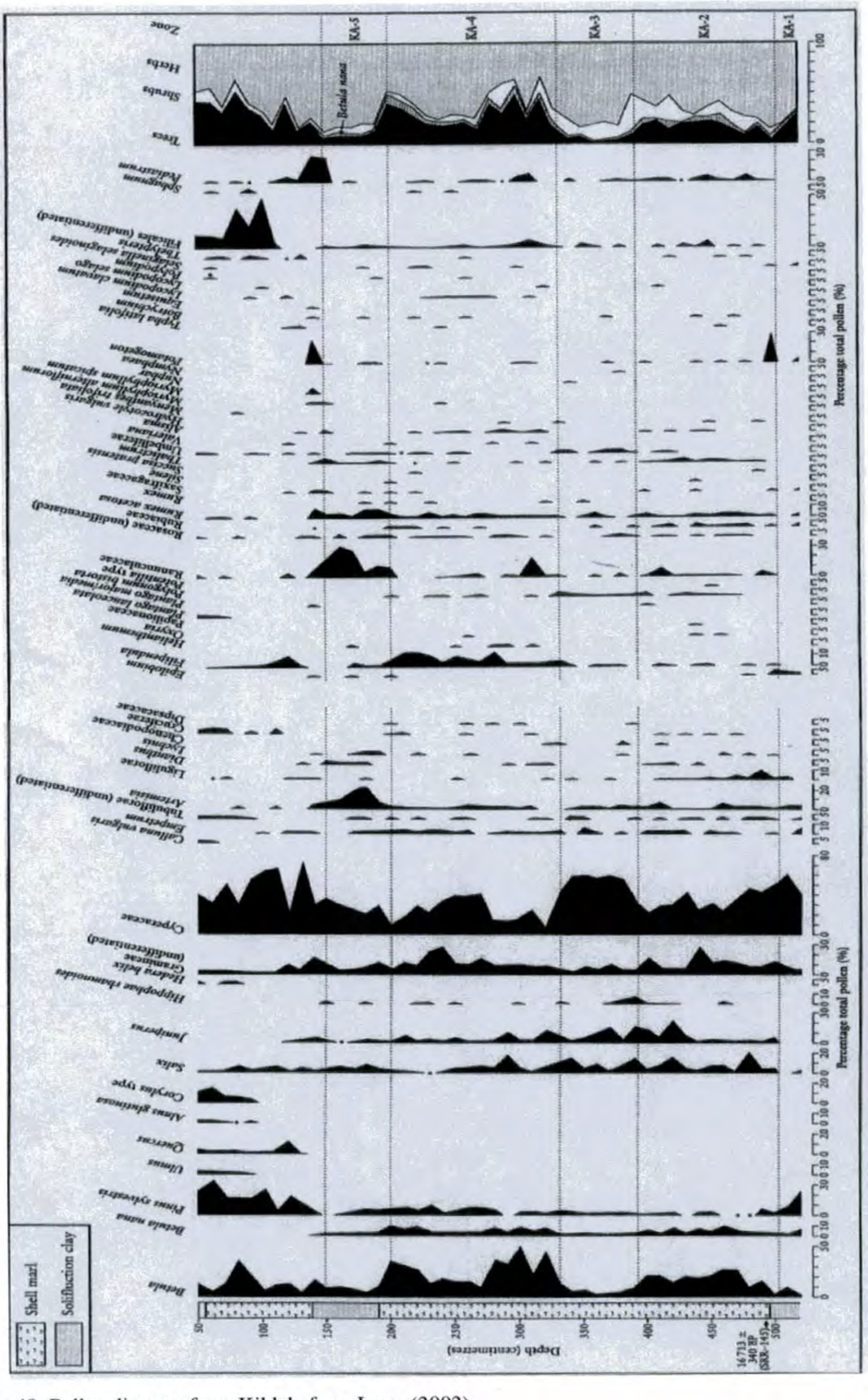


Figure 48: Pollen diagram from Kildale from Innes (2002)

6.3.2 The Tees Estuary

The proxy record taken from the Tees Estuary is a series of varves that were correlated to the GISP ice core record (Figure 49) (Plater *et al.* 2000). The thickness of each varve was used as a proxy for temperature. During warmer periods of time the annual melting of the ice mass increased. This resulted in an increase in the amount of sedimentation taking place within the lake resulting in a thicker rhythmite. However, it was stressed that varve thickness is unlikely to create a true climate proxy. Due to the thickness of the entire sequence it was proposed that deposition may have been affected by the proximity of the core to the lake shore. Thus the varve sequence may contain deltaic sediments which have not been influenced by climate. Therefore, the varve sequence was correlated with the GISP ice core record to verify the extent to which it could be used as a climate proxy. A good correlation was shown between the period of 18 000 and 16 000 cal. years BP suggesting that climate controlled sediment supply to the lake during this time (Plater *et al.* 2000). This proxy record shows an unstable cool climate during the period 16 000–17 750 cal. years BP. A sharp deterioration in climate is seen at 16 500 cal. years BP, matching the Kildale record. The varve record does not continue following this event. Although, the lithostratigraphical progression to Holocene deposits suggests that temperatures subsequently rose causing the ice to retreat and the lake to drain before sedimentation began again during the Holocene.

It is inferred that this laminated lake sequence must have taken place before deglaciation as if significant retreat had occurred prior to 16 000 cal. years BP relative sea level change would have inundated the lake basin (Shennan *et al.* 2006). Therefore, as Lake Tees was in existence at least between 18 000 and 16 000 cal. years BP ice must have been impounding it to the east by the North Sea Ice Lobe and west by ice from the Stainmore Gap. The lake also implies that ice flowing through the Stainmore Gap did not reach the east coast while it was in existence, as was proposed in Madgett and Catt (1978) and Foster (1986).

While this record was approximately dated using the correlation of the varve record with the GISP record absolute dating is problematic. A radiocarbon date taken from the upper section of the core indicated that the laminated sediments were deposited prior to 7173-7431 cal. years BP. In addition, a luminescence date of 18 365 ± 10 015 years BP was taken from the core (Plater *et al.* 2000). This high error term, which is thought to be a result of water content, makes the date highly unreliable.

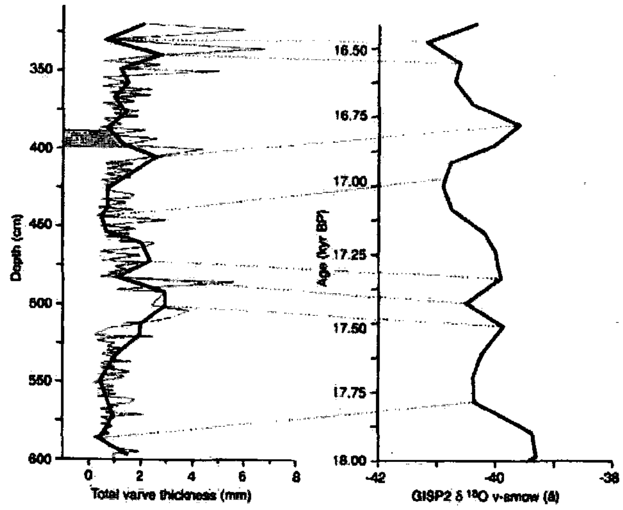


Figure 49: Comparison between varve thickness in Tees Estuary core T9 and the GISP ice core record from Plater *et al.* (2000). Shaded area signifies area fine grained luminescence dated to 18 365 ± 10 015 years BP.

6.3.3 Gransmoor

Proxy records from Gransmoor have also been used to investigate climate change in Yorkshire. Lowe *et al.* (1995) used a coleoptera record which had been correlated with the GISP ice core record. While this record closely matches the GISP ice core record it does not date back far enough to incorporate H1. Although a 4-5°C cooling is shown between 14 700 and 14 000 cal. years BP following an abrupt warming (Figure 47). Nevertheless, due to the dates marking the deglaciation in Yorkshire this climatic cooling does not seem likely to have been related to either ice advance at Upgang.

6.3.4 Climate of North Yorkshire during the last deglacial.

These three records show that climate in Yorkshire was highly variable around the time of the last deglaciation. Overall climate was slowly warming punctuated with cold spells that can tentatively be correlated with H1, 16 117 to 17 617 cal. years BP (Bond *et al.* 1992) and the LLS between 10 000 and 11 000 cal. years BP (Lowe and Walker 1997).

However, many factors aside from climate may have caused the ice lobe to readvance (E.g. internal ice dynamics). Most authors consider the tills deposited along the Yorkshire coast to have been deposited before 20 000 cal. years BP, with the Kildale date cited as a date indicating deglaciation of the whole area, although this is considered to be artificially old. In addition, the climate proxy records do not date far enough back to show a substantial amelioration in climate before this date that would have caused the ice sheet to retreat after the initial advance. The records also lack clarity and strong dating control at the beginning of the last deglaciation.

It should also be noted that the ice in Yorkshire was sourced entirely from outside the region (Harrison 1895) so that climate changes in the areas of ice accumulation would be significant. For example, a strong deterioration in climate in Scotland combined with an increase in precipitation would cause ice to readvance, although climate may not have changed significantly in Yorkshire. Therefore, it is difficult to tie early ice sheet readvances to the climatic record in Yorkshire. Nevertheless, Mayle *et al.* (1999) showed similar climatic patterns between Gransmoor (North Yorkshire), Whitrig Bog (Southern Scotland), and the GRIP ice core record (Figure 47). This similarity between a British climate record and the GRIP ice core record was also shown in Jones *et al.* (2002) at Hawes Water in Cumbria, (Figure 47). These inferred broadly similar climatic patterns throughout Great Britain. However, these records do not extent back as far as H1 as the areas investigated were still covered by ice. Although, if ice was readvancing into the area then it follows that temperature would drop in Yorkshire due to the presence of ice and associated albedo affect creating a microclimate.

6.4 Can a chronostratigraphic framework be established for Upgang?

The published data from the Tees Estuary along with the sediments from Dimlington provide an opportunity to correlate Upgang to these dated sites and, hence, establish a chronostratigraphic framework. If a correlation can be drawn with the tills at Dimlington or the Tees Estuary cores then the proposals relating to the glacial history of the area can be given greater strength and a relative chronology can be established.

6.4.1 Dimlington – Clast lithology

Both tills found at Dimlington overlie the Dimlington silts that have been dated to 20 931 – 22 281 cal. years BP (Penny *et al.* 1969). If the tills at Upgang can be correlated with these glacial sediments then it can be inferred that they were deposited as part of the same ice advance and, hence, are contemporaneous. One method of correlating the two sites is by clast lithology. If both sites have a similar erratic suite then it can be inferred that they have originated from the same source area and have similar ice flow direction histories.

A clast lithological sample was taken from both the lower, Skipsea, and upper, Withernsea, tills at Dimlington in order to compare them with the multiple samples from Upgang (Table 6). These two tills exhibit similar compositions to the tills at Upgang, although the percentages of each lithology vary.

Table 6: Table showing the percentage of each lithology counted depending on sedimentary LFA. Upgang values are averages from sites within the LFA.

Geology/Site	% D1	%D2	%D2A	%D2B	%D3	%SG	Upper Dimlington	Lower Dimlington
Limestone (Magnesium)		30.26	28.09	33.95	45.74	44.29	34.63	59.87
Shale	55.8	35.72	41.51	28.31	11.15	2.9	11.65	0.65
Sandstone (Quartz)		7.48	6.56	10.3	11.86	33.16	7.12	1.29
Mudrock	43.87	0.17	0.19	0.11		0.06	0.32	
Sandstone (Lithic)		6.87	6.66	8.03	5.97	1.01	9.06	1.29
Limestone (Carboniferous)	0.33	3.41	3.47	3.8	7.39	2.43	2.27	1.62
Greywacke		4.53	4.05	4.99	3.34	2.66	6.47	1.62
Quartzite		2.71	1.83	2.28	1.99	3.08	3.56	3.24
Quartz		1.44	0.77	1.74	1.21	3.43	0.97	0.65
Siltstone		1.53	1.25	1.74	2.34	0.3	3.56	
Wacke		1.47	1.93	1.19	0.64	0.18	6.15	0.97
Coal		0.46	0.29	0.87	2.2	0.41	4.53	
Dolerite		0.72	0.39	0.98	2.06	1.78	0.65	
Ironstone		0.75	0.39	0.33	0.92	0.53	0.32	0.32
Porphyry		0.43	0.29	0.22	0.64	1.84	1.29	
Granite		0.26	0.1			0.12	0.32	0.32
Andesite		0.32	0.39		0.64	0.47	0.32	0.32
Old Red Sandstone		0.2	0.29		0.21	0.06	1.94	
Rhyolite		0.29	0.19	0.22	0.5	0.3		0.65
Oosparite		0.23	0.58					
Sandstone (Arkose)		0.12	0.1	0.33	0.07	0.24	0.65	
Flint		0.17	0.29	0.11	0.14	0.24	1.62	6.47
Diorite		0.09	0.1	0.11				
Chalk					0.14		2.59	20.39
Pink Rhyolite		0.03			0.5	0.24		
Breccia		0.03		0.11	0.07			0.33
Porphritic Rhyolite		0.06	0.1					
Pyrite						0.06		
Phosphate		0.12	0.19	0.22		0.06		0.32
Hematite		0.03						
Chert		0.03		0.11		0.12		
Microgranite		0.06			0.28	0.06		

Many difficulties arise when comparing sites by lithological content because the majority of clasts are locally derived (Boulton 1996). Therefore, the dominant lithologies at Upgang are sourced from the underlying Jurassic bedrock, shale, mudrock, limestone and sandstone, while the majority of clasts at Dimlington are derived from the Cretaceous bedrock, chalk, flint and limestone. Consequently, it is the erratic suite that needs to be compared in order to establish if the sediments at Dimlington can be correlated with Upgang. This takes two courses, firstly looking at erratics from the samples at Dimlington that may have been derived from the Jurassic bedrock of Northern Yorkshire and secondly, the further travelled mainly igneous, erratics.

Both tills at Dimlington contain indicator erratics from the Jurassic sequence underlying Upgang suggesting that the ice depositing the tills at Dimlington had overrun northern Yorkshire. While many of the lithologies found at Dimlington could have been derived from the Jurassic sequence to the north, for example sandstone, they commonly belong to a variety of geological formations and, hence, will not be used for correlation. Nevertheless shale, Jurassic mudstone and Jurassic ironstone all distinctly belong to the Jurassic formation of North Yorkshire. However, this bedrock sequence extends offshore for over 40 miles, so the ice that deposited the tills at Dimlington has not necessarily directly overrun the Upgang area but has still flowed from north to south.

Dimlington also has a similar far travelled erratic suite to Upgang, suggesting that one ice lobe deposited the sediments at both sites. The upper, Withernsea Till from Dimlington shows relatively high quantities of coal and Whin Sill dolerite from the north east of England along with porphyry, granite and andesite from the Cheviot Hills. The lower, Skipsea Till does not contain the erratics from north east England but does exhibit rhyolite, andesite and granite from the Cheviots.

Tentatively, this suggests that both tills at Dimlington were deposited by the same North Sea Ice Lobe that deposited the tills at Upgang. However, it should be noted that none of these igneous erratics in either till are found in any significant quantity, although this is to be expected as the number of distally derived clasts decreases with transport length (Larsen and Mooers 2004). Also, as only one sample has been taken from each till it is possible that other indicator erratics may be present at the site but have not been collected due to the small sample size. While other lithographical studies have been carried out on the Yorkshire coast, the original clast counts and quantity of lithologies found have not been published in detail preventing the expansion of the data set in this manor. In addition, the robustness of lithostratigraphical correlations has been questioned. In north Norfolk, Hamblin *et al.* (2000) suggested that the Happisburgh Diamict represented an oxygen isotope stage (OIS) 16 event following the separation of three tills according to their lithological suites. This was later developed by Lee *et al.* (2004) who correlated other tills with this event based on the correlation of lithological suites. However, Banham *et al.* (2001) queried this initial proposal as an 'OIS reading of the facts' (Banham *et al.* 2001 p 8). It is stated that aside from the lithological content that the tills are very similar. In addition, assigning the Happisburgh Diamict to an OIS 16 event was considered to be in direct conflict with biological and lithostratigraphical evidence. Thus, correlation and differentiation on a purely lithological basis can be contentious. Nevertheless, many of the correlations between the tills along the Yorkshire coast have been based on the far-travelled proportion of clast lithological counts.

6.4.2 Tees Estuary - Geochemistry

Geochemical data taken from Plater *et al.* (2000) provides another opportunity for correlation between Upgang and a dated site. Geochemical analysis was carried out on several cores taken from the Tees Estuary. For the purpose of this study only the geochemical results taken from the tills found at the base of the core have been used. This data is an average from these tills opposed to elemental data from each individual sample. The range of elements analysed in the Tees Estuary is smaller than at Upgang. Therefore, only the elements in both data sets have been used for correlation by cluster analysis (Figure 50) and chi-squared testing (Figure 51). Those elements that have only been measured at one location have been removed in order to gain a stronger correlation.

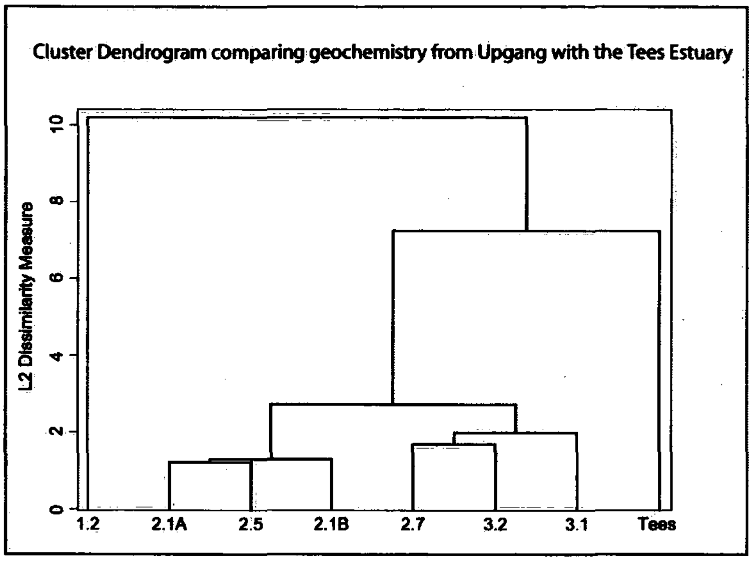


Figure 50: Cluster analysis dendrogram comparing the diamict from Upgang with the till found in the Tees Estuary.

	x^2	Reduced x^2
Tees vs D2	463.63	92.73
Tees vs. D3	283.22	94.41

Figure 51: Chi-squared values correlating the geochemistry from the Tees Estuary till (Plater *et al.* 2000) and that at Upgang.

Neither of these methods displays a strong correlation, with very high reduced chi-squared values and high dissimilarity values with both D1 and D2. However, correlation between sites using geochemistry has many of the same difficulties as clast lithologies. Locally derived elements make up the dominant proportion of a geochemical data set (Larsen and Mooers 2004). Therefore, as the Tees Estuary overlies Triassic bedrock with Jurassic bedrock below Upgang, the geochemistry of the tills in both areas will be distinctly different. Other factors will also influence the geochemistry data leading to poor correlation, for example weathering. As the tills sampled from the Tees Estuary have been obtained from estuarine cores they could have been weathered differently from those taken at the exposed coastal section of Upgang. Problems also arise when comparing samples using a data set derived and published by another author, especially if their techniques are unknown (Banham *et al.* 2001). Elemental analysis at each site was carried out using different equipment which could have different sensitivity levels. Also the data from the Tees looked at a smaller set of elements than the samples from Upgang. While only the elements analysed at both sites where used in the correlation statistics the values from Upgang could be affected as they were quantified initially as part of a larger data set.

Nevertheless, this does not mean that the ice depositing the sediments at Upgang didn't overrun the Tees Estuary area as the red colouring of the upper till at Upgang, D3, was attributed to the Triassic Marls underlying Middlesbrough and extending offshore (Radge 1939) (Figure 4). This correlation in colour was further developed by Agar (1954) suggesting that the Tees Red and Drab tills correlated with the Purple and Drab clays at Upgang and along Holderness based on colour. It should be noted, however, that correlation on colour alone is inconclusive as many other factors can alter colour, for example weathering profiles (Eyles and Sladen 1981).

6.5 Timing of the ice sheet advances at Upgang.

It is necessary to establish the timing of the advance-retreat-readvance signal at Upgang so that any correlations with Heinrich Event 1 (H1) and other British Last Glacial Maximum (LGM) readvances can be made. The tentative correlation between the tills at Upgang and Dimlington discussed in Section 6.4, along with deglaciation dates from Yorkshire, allow a relative chronology to be discussed for this glacial sequence.

The date of 20 931-22 281 cal. years BP (Penny *et al.* 1969) from Dimlington was taken from a layer of silt underlying both the Skipsea and Withernsea tills. As a result both of the Dimlington tills must have been deposited after this date, with at least the Skipsea Till representing the LGM along the Holderness coast. The correlation between the tills at Dimlington and Upgang implies that both Upgang tills were also deposited after 20 931 – 22 281 cal. years BP. Deglacial dates for Yorkshire come from the vegetation records of Kildale, 19 144-20 537 cal. year BP (Jones 1977), and Roos Bog, 14 360-12 628 cal. years BP (Beckett 1981). While the Kildale date was believed to be artificially old due to hard water errors (Innes 2002), the date from Roos Bog was thought to be too young. Nevertheless, these dates show that Yorkshire was deglaciated at some time between 19 000 and 14 000 cal. years BP. These dates along with the date from Dimlington show that the advance-retreat-readvance signal at Upgang must have been deposited between 21 000 and 19 000-14 000 cal. years BP.

Evidence from the climatic records discussed in Section 6.3 show a clear cooling phase between c. 16 000 cal. years BP and 17 000 cal. years BP suggesting that the readvance signal could have been driven by climate. This is supported by the strong association between the British climatic records and the Greenland ice core records which show clear drops in temperature associated with H1. Therefore, tentatively the readvance sediments at Upgang can be correlated to H1. In addition, the existence of Lake Tees around this time implies that ice must have been present at least as far south as the Tees during this period. However, the readvance signal represented by D3 may be a result of internal ice dynamics and not climate.

Thus, it is proposed that a major ice advance occurred after 21 000 cal. years BP depositing the lower tills, D1 and D2 at Upgang extending at least as far south as Dimlington. It is therefore, these tills that represent the LGM event at Upgang. The ice lobe then retreated further north than Upgang resulting in the deposition of the proglacial lake sediments, SG. Ice then readvanced to a location south of Upgang depositing D3. It is this readvance that is cautiously correlated to climatic cooling and H1. Without absolute dating control at Upgang this hypothesis cannot be strengthened.

6.6 How does Upgang fit in with proposed Last Glacial Maximum (LGM) readvance signals in Great Britain and Northern Ireland?

A number of glacial readvances at the end of the last glacial maximum (LGM) have been proposed for different parts of Great Britain and Ireland, on the basis of sedimentological and chronological evidence. The two most significant areas of readvance in terms of dating control and the number of papers published are in the Irish Sea Basin (ISB) (McCabe *et al.* 1998, Merritt and Auton 2000, Thomas *et al.* 2004, McCabe and Clark 2003, Roberts *et al.* 2006, McCabe *et al.* 2007b) and in the east of Scotland (Hall and Jarvis 1989, Peacock 1997, McCabe *et al.* 2007a). Dates relating to these sites along with other sites of readvances and the key dates from Yorkshire are shown in Table 7 and Figure 52.

Table 7: Published dates marking the deglacial chronology of the east coast of Britain.

Site	Radiocarbon Date	Calibrated Date	Site details	Literature Source
Heinrich Event 1	13 490 – 14 500	16 117 – 17617*	Not specified	Bond *et al.* (1992)
Bremanger Moraine, Norway	15 000 – 13 300	18 085 – 18556 * 15448 – 16138 *	Various microfossils from glaciomarine unit above till.	Nygard *et al.* (2004)
Killard Point, Northern Ireland	13 785 ± 115 13 955 ± 105	c. 16 500	E. clavatum from marine mud in interbedded outwash from Killard Point moraine.	McCabe *et al.* (1998)
Jurby Head, Isle of Man	15 150 ± 350 12 890 ± 360	18 578 - 17 624 15 992 - 14 370	Organic mud overlying the Jurby Formation.	Thomas *et al.* (2004)
Jurby, Isle of Man	12 276 ± 45	13 983 – 14 402	Salix within kettle hole.	Roberts *et al.* (2006)
Glen Balleria, Isle of Man	12 492 ± 92	14 170 – 14 996	Salix within kettle hole.	Roberts *et al.* (2006)
Blenham Bog, Cumbria	14 330 ± 230 14 280	16 456 – 17 985 * 16 675 – 17 428 *	Lowest organic material in kettle hole.	Pennington and Bonny (1970)
St Bees, Cumbria	N/A	15 500 – 14 300	Not specified.	Thomas *et al.* (2004)
Dundalk Bay, Northern Ireland	14 157 ± 69	16 455 – 17 303 *	E. clavatum from deformed mud below ice pushed morainic sediments.	McCabe *et al.* (2005)
Wester Ross, Scotland (Cosmogenic date)	N/A	15 500 ± 240 (^{10}Be) 17 900 ± 310 (^{10}Be)	Boulders within the Wester Ross moraines.	Everest *et al.* (2006)
St Kilda, west Scotland	15 200	18 545 – 18 722 *	From vibrocores south of St Kilda.	Peacock *et al.* (1992)
St Fergus, east Scotland	14 915 ± 210	17 386 – 18 727 *	St Fergus Silts, towards top of deposit.	Hall and Jarvis (1989)
Lunan Bay, east Scotland	17 065 ± 50 17 720 ± 50	20 250 ± 40 21 130 ± 90	From raised marine mud overlain by ice contact gravel.	McCabe *et al.* (2007)
Gallowflat, east Scotland	13 655 ± 45 13 675 ± 40	16 355 ± 130 16 385 ± 125	From laminated muds 1m above glacial diamict	McCabe *et al.* (2007)
Dimlington, Yorkshire	18 240 ± 250	20 931 – 22 281 *	Moss within the Dimlington silts.	Penny *et al.* (1969)
Dimlington, Yorkshire (TL date)	N/A	17 500 ± 1.6 x 10^3 (TL)	The loess component of a solifluction deposit overlain by till.	Wintle and Catt (1985)
Kildale, Yorkshire	16 713 ± 340	19 144 – 20 537 *	Moss within a thick shell marl. (date area was ice free)	Jones (1977)
The Bog, Roos, Yorkshire	13 045 ± 270	14 360 – 12 628*	From a kettle hole within Withernsea till.	Beckett *et al.* (1981)

* Calibrated from published 14C date using Calib Rev 5.0.1 (two sigma ranges)

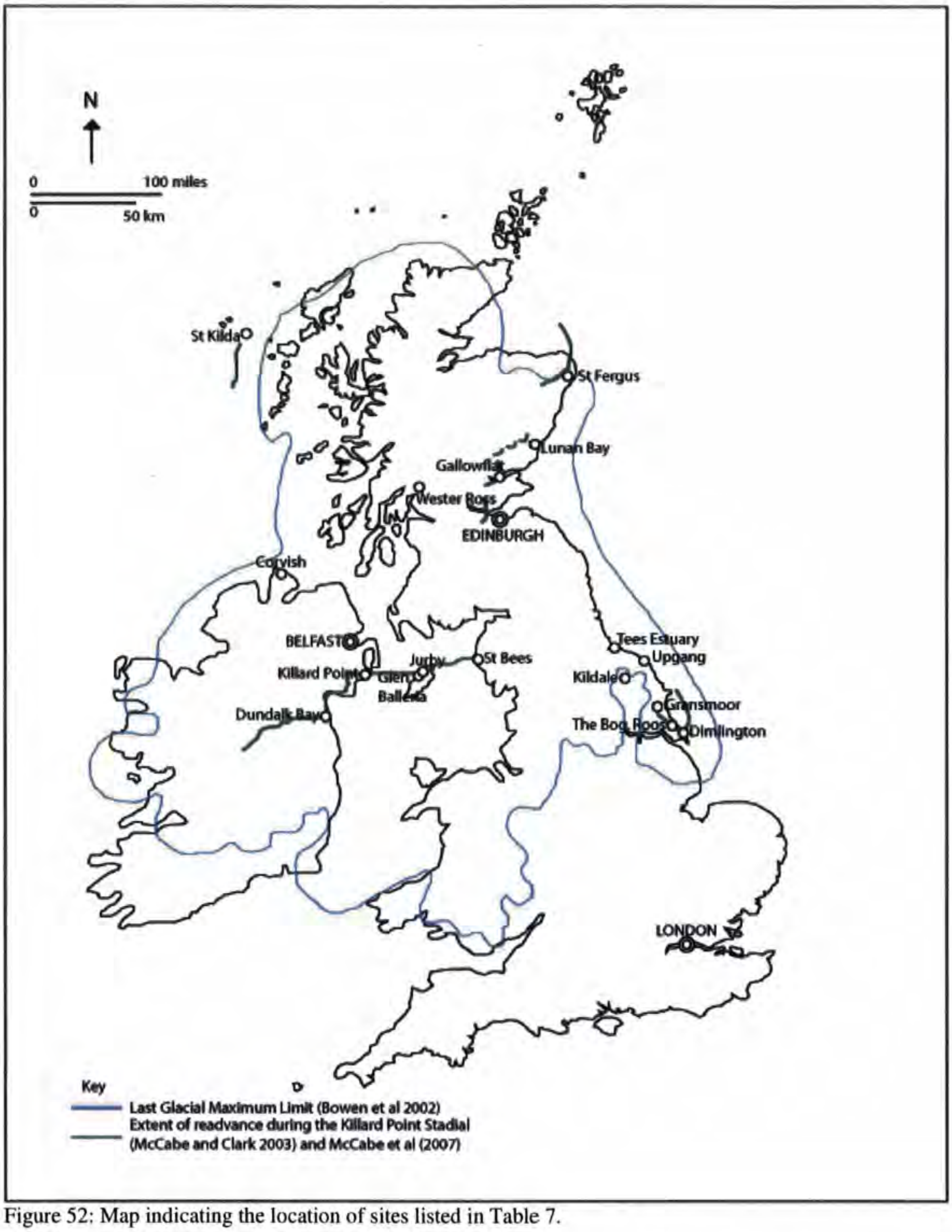


Figure 52: Map indicating the location of sites listed in Table 7.

The lack of absolute dating control on the site at Upgang makes it impossible to correlate the readvance studied with other readvances around Great Britain and Northern Ireland with any great certainty, especially with regards to H1. However, relationships have been drawn due to the dates available within Yorkshire. As a result of the dates underlying the till at Dimlington, 20 931 – 22 281 cal. years BP, the overlying glacial sediments must have been deposited after this date. As discussed in Section 6.5 the evidence from climatic records allows the tentative correlation of the readvance till, D3, at Upgang to H1. As a result the readvance signal at Upgang would correspond with the late glacial readvances in the ISB, east Scotland and other parts of Great Britain. However, without absolute dating of the sediments at Upgang no stronger hypothesis can be formed regarding the chronology of events and the potential links with other sites around the country.

6.7 Project limitations and further work.

The major limitation of this project is the lack of direct dating control on the sediments at Upgang. This prevents any strong conclusion regarding the timing of the readvance being formed. However, a readvance signal was established from the sedimentological analysis and it has been possible to consider the site within the context of the glacial history of the Yorkshire coast. A corollary of this limitation is the inability to strongly correlate Upgang to any of the dated sites in Yorkshire. This is a result of the limitations attached to lithostratigraphical and geochemical correlation along with time constraints; preventing geochemical analysis of sediments at Kildale, addition geochemical sampling from the Tees Estuary and a greater clast lithological analysis of Dimlington. Strong correlation between sites would have given the site a clearer, stronger chronology in the absence of direct dating evidence.

In order to overcome this limitation it may be possible to optically stimulated luminescence (OSL) date the sand units within SG. This would help to provide a clearer chronology of the sediments at Upgang. In addition, it would be an extra date relating to the glacial history of Yorkshire and would allow correlation with other deglacial stratigraphies around Great Britain and Ireland.

Additional further work would involve the study of other sites north of Holderness to assess the extent of the proposed readvance. These sites along with those in Holderness also need to be studied to establish if there is evidence for only one ice lobe advance in the area. Detailed clast lithology studies would be vital in this context in order to verify the presence of only a northern North Sea Ice Lobe.

7. Conclusions

The coastal section of Upgang, North Yorkshire, shows a distinct advance - retreat-readvance sequence. This consists of a grey diamict, D1, at the base and an overlying brown diamict, D2. These tills were deposited during the initial advance of ice. A sand and gravel lithofacies association was then deposited during the retreat of the ice. The section is topped by a reddish brown diamict, D3, which represents the readvance event.

D1 is either a glaciotectonite or an overrun proglacial thrust feature made up from the local shale and mudrock bedrock. Deposited during or prior to the initial ice advance the LFA was detached from the bedrock, transported a short distance, broken up and then deposited. This LFA only outcrops in parts of the cliff section with the degree of brecciation varying between highly deformed and relatively untouched.

D2 directly overlays D1 with a sharp distinct boundary between the two. D2 is a homogenous matrix supported diamict which has been interpreted as a subglacial traction till dominated by deformation processes. The clast form and clast fabric data along with the presence of boudins, folds, rafts and laminations support this hypothesis. In addition, the sharp boundary between D1 and D2 is interpreted as a décollement plane which is can also be seen as indicative of deformation processes occurring at the bed of a glacier. Till has been thickened during deposition by till stacking. Clast lithological analysis from this LFA indicates that ice was sourced from Southern Scotland flowing through the Cheviot Hills and the North East of England.

Ice then retreated from the Upgang area resulting in the deposition of the clays, silts, sands and gravels. This LFA represents a lake that was subsequently infilled through proglacial subaerial sandur sedimentation. At the base of the sequence SG1 represents a deep lake with fines and sands deposited through suspension. SG2 then represents the shallowing and infilling of this lake with laminated and rippled sands deposited by underflow and suspension deposition. SG3 is a gravel facies, with coarsening upward units and trough bedded channels indicating proximal channels and indicates an increasing proximity to the ice front. The presence of SG4, a thin layer of sand, indicates either proglacial or subglacial sheet flow within a wide shallow channel.

This ice readvance deposited the reddish brown diamict of D3. D3 has also been interpreted as a subglacial traction till dominated by deformation processes. As with D2 this is due to the clast form and clast fabric data supported by boundinaged sand and clay inclusions. The LFA was thickened by the process of till stacking resulting from Upgang's close proximity to the high ground around the North Yorkshire Moors. Clast lithology shows that D3 has the same provenance as D2, with ice originating from Southern Scotland and Northern England.

This advance – retreat – readvance signal does not mirror the published stratigraphy of the Yorkshire coast. It is proposed that the section at Upgang has been deposited by a single northern ice lobe, and not a multilayered Lake District and northern ice mass as previously suggested. This questions the modes of deposition proposed for the Skipsea and Withernsea Tills along the remainder of the Yorkshire coast. The extent of the ice readvance seen at Upgang is unknown and further work on the exposures south of this area is needed to verify this limit.

Clast lithological studies at Upgang and Dimlington show that both sites have a similar far travelled erratic suite. Hence, it can be hypothesised that the two tills at Upgang were also deposited post 21 000 cal. years BP by a single northern ice lobe. It is proposed that the initial advance depositing D1 and D2 occurred immediately after c.21 000 cal. years BP representing the LGM till at Upgang.

The climate records from Kildale and the Tees Estuary show a decline in climate around 16 000 cal. years BP, broadly corresponding to Heinrich Event 1. Therefore, this deterioration in climate is cautiously proposed as the mechanism causing the ice readvance that deposited D3 at Upgang. Thus, the readvance signal is tentatively correlated with Heinrich Event 1 and the other Last Glacial Maximum readvances around Great Britain.

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Appendix 1: Clast lithological table

Geology/Site	D1	D2	1.1	1.2	1.3	2A.1	2B.1	2A.2	2B.2	2.3	2A.4	2B.4	2.5	2.6	2.7
Limestone (Magnesium)	34.63	59.87				22.77	33.11	30.93	36.11	24.64	30.62	32.75	26.79	30.92	36.36
Shale	11.65	0.65	6.81	97.1	50.14	46.97	32.08	39.94	12.85	35.75	37.64	38.01	49.7	41.79	17.01
Sandstone (Quartz)	7.12	1.29				6.92	8.87	7.21	17.01	6.04	5.62	5.85	3.87	5.07	10.85
Mudrock	0.32		92.92	2.9	49	0.58				0.24		0.58	0.3		0.29
Sandstone (Lithic)	9.06	1.29				8.65	3.41	4.8	13.89	10.63	6.46	7.02	3.57	2.66	8.21
Limestone (Carboniferous)	2.27	1.62	0.27		0.86	2.02	4.44	6.31	4.51	2.9	2.25	2.63	2.38	3.62	3.52
Greywacke	6.47	1.62				4.03	5.12	1.5	5.56	5.56	6.46	4.39	3.87	3.62	5.28
Quartzite	3.56	3.24				0.86	2.39	1.2	3.13	3.86	3.37	1.46	1.79	4.59	3.81
Quartz	0.97	0.65				1.15	1.71	0.6	2.08	1.93	0.56	1.46	1.49	0.72	2.93
Siltstone	3.56					1.44	1.71	1.5	1.04		0.84	2.34	0.89	2.17	3.52
Wacke	6.15	0.97				1.15	1.71	3.3	2.08	2.9	1.4		1.49	0.48	0.29
Coal	4.53					0.58	0.34			0.24	0.28	2.05		0.24	0.88
Dolerite	0.65					0.86	2.39	0.3		1.69		0.58		0.72	0.59
Ironstone	0.32	0.32				0.58	0.68		0.35	1.21	0.56		0.6	0.97	2.35
Porphyry	1.29						0.34	0.6		0.24	0.28	0.29	0.3	0.72	1.47
Granite	0.32	0.32						0.3		0.48			1.19		0.59
Andesite	0.32	0.32				0.58				0.24	0.56		0.3	0.48	0.88
Old Red Sandstone	1.94							0.9					1.19		
Rhyolite		0.65					0.34		0.35	0.48	0.56			0.48	0.59
Oosparite						0.29		0.3		0.24	1.12			0.24	
Sandstone (Arkose)	0.65						0.68	0.3	0.35						
Flint	1.62	6.47				0.29	0.34				0.56		0.3		0.29
Diorite										0.24	0.28	0.29			
Chalk	2.59	20.39													
Pink Rhyolite															0.29
Breccia		0.33							0.35						
Microgranite														0.48	
Chert									0.35			0.29			
Porphritic Rhyolite										0.24	0.28				
Hematite										0.24					
Phosphate		0.32				0.29	0.34				0.28				
Pyrite															

Geology/Site	3.1	3.2	3.3	3.4	3.5	SG1	SG2	SG3	SG4
Limestone (Magnesium)	36.8	52.45	52.92	41.96	45.42	47.67	53	40.98	33.42
Shale	34.42	3.15	3.78	5.49	2.92	3.1	0.48	1.64	6.38
Sandstone (Quartz)	6.82	10.49	13.75	16.08	13.75	30.23	28.54	36.89	38.27
Mudrock							0.24		
Sandstone (Lithic)	5.04	4.9	6.53	5.49	8.33	1.16	0.24	1.09	1.53
Limestone (Carboniferous)	3.56	7.34	6.19	9.41	12.08	1.36	1.44	4.64	2.81
Greywacke	3.86	3.5	2.06	4.31	2.92	1.16	3.36	2.19	4.34
Quartzite	0.3	1.75	1.37	3.53	3.75	1.94	3.84	3.01	3.83
Quartz	1.19	1.4	1.37	0.78	1.25	3.88	3.84	2.46	3.32
Siltstone	1.48	3.5	1.72	2.75	2.5		0.24	1.09	
Wacke	0.3	1.05		0.39	1.67	0.58			
Coal	1.48	2.45	2.75	1.57	2.92	0.78		0.82	
Dolerite	2.08	1.05	3.09	3.14	0.83	2.52	1.2	1.37	1.79
Ironstone		2.8	0.69	0.39	0.83	0.39	0.24	1.64	
Porphyry	0.59		0.69	1.57	0.42	2.33	1.68	1.37	1.79
Granite						0.39			
Andesite	0.3	0.7	1.03	1.18		0.78	0.24		0.77
Old Red Sandstone		0.35	0.69			0.19			
Rhyolite	0.59	1.05	0.69			0.58	0.24		0.51
Oosparite									
Sandstone (Arkose)			0.34				0.24		0.77
Flint		0.7				0.19	0.24		0.51
Diorite									
Chalk		0.7							
Pink Rhyolite	0.59	0.35		1.57		0.39		0.27	
Breccia			0.34						
Microgranite	0.59			0.39	0.42		0.24		
Chert						0.19	0.24		
Porphritic Rhyolite									
Hematite									
Phosphate		0.35					0.24		
Pyrite						0.19			